STATE OF OHIO
DEPARTMENT OF NATURAL RESOURCES
DIVISION OF GEOLOGICAL SURVEY
Horace R. Collins, Chief

Report of Investigations No. 110

HYDRAULIC PROPERTIES OF A LIMESTONE-DOLOMITE AQUIFER NEAR MARION, NORTH-CENTRAL OHIO

by

Stanley E. Norris
U.S. Geological Survey

Columbus
1979

Ohio Department of Natural Resources
DIVISION OF GEOLOGICAL SURVEY
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CONTENTS

FIGURES

CONTENTS

TABLES

HYDRAULIC PROPERTIES OF A LIMESTONE-DOLOMITE AQUIFER NEAR MARION, NORTH-CENTRAL OHIO

by

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ABSTRACT

Ten exploratory wells, three for possible production and seven for observation, were drilled at a site near Marion in north-central Ohio, and long-term aquifer tests were made to determine the hydraulic properties of a regionally extensive carbonate-rock aquifer. The central question was the feasibility of seasonally pumping up to 3,000 gallons per minute (189 liters per second) of ground water. About 260 feet (79 meters) of Silurian limestone and dolomite deposits covered by about 31 feet (9 meters) of glacial till underlie the test site. The most permeable part of the carbonate-rock aquifer is 70 feet (21 meters) thick, lies between the depths of 80 and 150 feet (24 and 46 meters), and is characterized by fissures and cavities. Prior to pumping, the potentiometric surface in fully penetrating wells was 3 to 10 feet (1 to 3 meters) below land surface, slightly higher than the water level in shallow wells open in the top of the carbonate rock and slightly lower than the water table in the till.

The pumping of paired shallow and deep wells indicated a poor hydraulic connection between the high-yielding stratum and the overlying poorly productive part of the carbonate-rock aquifer. Preliminary testing, including specific-capacity tests, showed that hydraulic conductivity varies laterally as well as vertically, being highest near the center of the test-well array at well 2 and lowest 3,000 feet (914 meters) west of well 2 near well 4.

Two aquifer tests were made in late 1973 in which well 2 was pumped. In the first test, terminated after 9 days because of power failure, the pumping rate was 494 gallons per minute (31 liters per second); in the second test, lasting 90 days, the pumping rate was 369 gallons per minute (23 liters per second). Semilogarithmic distance-drawdown graphs for selected pumping periods indicate that the shape of the cone of depression became essentially stable after less than 2 days of pumping. The region of locally high hydraulic conductivity was reflected by the flatness of the graphs in the vicinity of the pumped well.

Transmissivity calculated from time-drawdown data by the straight-line semilogarithmic method averaged about 3,600 square feet per day (330 square meters per day); the storage coefficient averaged about 10^{-4}. The graphs showed the effect of recharge by leakage from the glacial till after 4,000 to 6,000 minutes of pumping.

The leakance factor, k'/m', where k' is the vertical hydraulic conductivity of the till and m' is its thickness, determined by the steady-state equation of Jacob was 3.2×10^{-6} per day for drawdown data collected after 20 days of pumping and 2.9×10^{-6} per day for data collected after 25 days of pumping. The leakance factor calculated by relating the quantity of water pumped to the volume of the cone of depression was 9.6×10^{-6} per day and 9.8×10^{-6} per day, respectively, for the same pumping periods. The hydraulic conductivity of the glacial till based on an average regional saturated thickness of 35 feet (11 meters) is in the range of 1.0×10^{-4} to 3.4×10^{-4} feet per day (3.0×10^{-5} to 1.0×10^{-4} meters per day).

The seasonal pumping of approximately 3,000 gallons per minute (189 liters per second) of ground water desired for a wildfowl preserve is not feasible based on the aquifer tests. It is estimated that about half the required quantity could be obtained from the three existing production wells; however, such withdrawal would be accompanied by significant drawdown over a wide area and might adversely affect the water supply of many residents.

Despite the fact that the carbonate-rock aquifer is neither homogeneous nor isotropic, the test results probably can be applied to other areas in western Ohio with a fair degree of confidence. The range of hydraulic conductivity values determined for the glacial till also may have useful transfer value.

INTRODUCTION

This report describes results of aquifer tests (including one of 90-day duration) made between May 1972 and February 1974 to determine the hydraulic characteristics of a carbonate-rock (chiefly dolomite) aquifer overlain by glacial till at a site 7 mi (11 km)[1] west of Marion, Marion County, in north-central Ohio (fig. 1). The test site, herein referred to as the Big Island site (from the crossroads hamlet 4 mi (6 km) to the northeast), is within the Big Island Wildlife Area, a 1,900-acre (770-ha) tract of state-owned land used as a public hunting area. The Ohio Department of Natural Resources, Divisions of Water, Geological Survey, and Wildlife, made the investigation jointly and in cooperation with the U.S. Geological Survey.

The opportunity to make an extended test of this carbonate-rock aquifer, which is an important source of water in much of western and central Ohio, was an outgrowth of state water-plan studies in southwestern and central Ohio. Ten exploratory wells were drilled at the Big Island site. Three wells were 10 inches (250 mm) in diameter and constructed for pumping and the remaining wells were 6 inches (150 mm) in diameter and constructed for observation.

One of the objectives was to collect and interpret aquifer-test data from fully and partially penetrating wells in

[1] Conversion factors for the terms used in this report are listed below:

Multiply English unit	*by*	*to obtain metric unit*
inches (in)	25.4	millimeters (mm)
feet (ft)	0.3048	meters (m)
miles (mi)	1.609	kilometers (km)
feet per mile (ft/mi)	0.19	meters per kilometer (m/km)
square feet (ft^2)	0.0929	square meters (m^2)
cubic feet (ft^3)	0.02832	cubic meters (m^3)
acres	0.4047	hectares (ha)
gallons (gal)	3.785	liters (l)
gallons per minute (gal/min)	0.06309	liters per second (l/s)

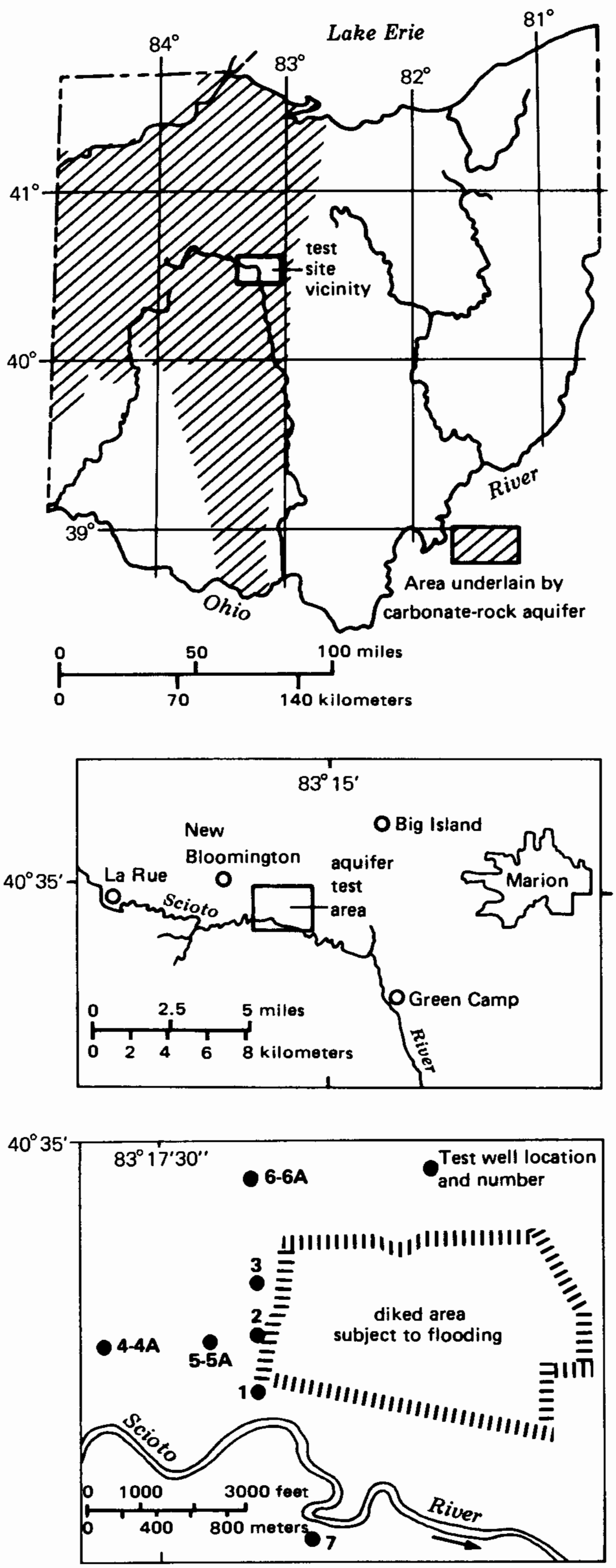

FIGURE 1.–Area in Ohio underlain by the carbonate-rock aquifer, aquifer test area, and location of test wells.

the expectation that these data and their interpretation would have important transfer value to more heavily pumped areas of the state. Carbonate-rock aquifers rarely meet the assumptions, such as isotropy, on which many commonly used aquifer-test equations are based. A key question partly answered in this investigation relates to the validity of short-term aquifer tests of 8- to 48-hour duration in predicting long-term pumping effects in carbonate-rock aquifers.

Of more immediate interest to officials of the Ohio Division of Wildlife was the degree to which the aquifer might be used without adverse effects on farm and domestic wells in the surrounding area for seasonal large-scale pumping to flood parts of the wildlife area, creating a temporary habitat for migratory wildfowl. It was hoped that ground water could be used, altogether or in part, as a substitute for water now being pumped from the Scioto River for this purpose. Pumping from the river at rates of 3,000 to 3,500 gal/min (189 to 221 l/s) in late summer and fall, when the pumpage may amount to a third or more of the total flow, is of concern to officials of the City of Marion because the municipal water-supply intake is a few miles downstream from the Big Island Wildlife Area. The Ohio Division of Wildlife, which annually undertakes the task of artificially inundating about 380 acres (150 ha) of this flat poorly drained land with an average of 3 to 4 inches (80 to 100 mm) of water, contributed materially to this investigation by paying for the construction of a pumphouse and installation of an electrically powered turbine pump.

The author acknowledges the work of J. J. Schmidt, hydrologist with the Ohio Division of Water, and H. B. Eagon, Jr., hydrologist, formerly with the Ohio Division of Water, for designing the test pattern and supervising the drilling of the wells for this investigation. In 1972 Mr. Eagon made or supervised tests of approximately 24-hour duration at two of the three pumped wells and made the long-term tests of well 2 with the assistance of personnel of the U.S. Geological Survey. The data from the well 2 tests largely constitute the subject of this report.

Appreciation is expressed to Emery Jividen and other officials of the Ohio Division of Wildlife for assistance and cooperation in the field work and to L. J. Harstine, Ohio Division of Water, and D. E. Johe, Jr., Ohio Environmental Protection Agency, for helping collect the data.

THE AQUIFER SYSTEM

Underlying the till plain in most of western Ohio are limestone and dolomite beds of the Silurian Lockport Dolomite and Bass Islands Group[2] (chiefly dolomite), which together or individually constitute an important regional aquifer (Norris and Fidler, 1971, 1973).

The attitude of the carbonate rocks in western Ohio is controlled by the broad, low Cincinnati arch, a regional structural feature whose long axis trends almost north-south between Toledo and Cincinnati. The beds are nearly horizontal on the top of the arch and dip away from the

[2]The usage of the term Bass Islands Group for rocks overlying the Lockport Dolomite in the study area does not agree with current terminology of the Ohio Division of Geological Survey. See Ohio Geological Survey Report of Investigations 100, *Silurian rocks in the subsurface of northwestern Ohio*, by Adriaan Janssens, 1977.

crest to the west, north, and east at low angles. The Big Island test site is on the east side of the arch about 15 mi (24 km) from the crest. The regional dip of the arch near the test site is about 7 ft/mi (1 m/km) to the east.

The Lockport Dolomite, 132 ft (40 m) thick at the test site, is finely to coarsely crystalline light-brown to gray dolomite. Where the unit is exposed, it occurs typically in beds ranging from 1 to 5 ft (0.3 to 2 m) thick, locally grading into reeflike masses with little or no discernible bedding. The Lockport is a good source of water for farm and domestic wells in parts of western Ohio where it directly underlies the till and has been subject to weathering. The unit yields relatively little water to wells where it is deeply buried by younger rocks, as in the Big Island area.

The overlying Bass Islands Group is generally more permeable than the Lockport Dolomite and mostly consists of thin-bedded crystalline to granular argillaceous brown to drab dolomite, but there is much variation in lithology. An especially permeable Silurian unit is the Newburg zone, a discrete zone of high hydraulic conductivity, 10 to 15 ft (3 to 5 m) thick, at or near the base of the Bass Islands Group (Norris and Fidler, 1973, p. 36). The Newburg zone is present in much of western Ohio and is believed to extend into the oil and gas fields of the eastern part of the state, where it constitutes a deep-seated brine-yielding zone (Norris, 1956).

The most permeable part of the carbonate-rock aquifer at Big Island and the only part capable of yielding a significant quantity of water to wells is a 70-ft (21-m) section, which includes the Newburg zone, near the base of the Bass Islands Group. Caliper logs of the test wells (fig. 2) show the rocks in this interval to contain prominent cavities, a marked contrast to the rocks above and below. In drilling the test wells drillers reported little water above 80 ft (24 m) or below 150 ft (46 m), the approximate upper and lower boundaries of the cavity zone. The bottom of the cavity zone is 2 to 7 ft (1 to 2 m) above the Lockport Dolomite, and the top is about 50 ft (15 m) below the base of the till.

The caliper-log evidence and the drillers' reports indicating the depth and thickness of the productive zone were confirmed by current-meter flow tests. To make these flow tests a deep-well current meter was installed in a cylinder, 4 inches (100 mm) in diameter, which was fitted with a neoprene collar of the same nominal diameter as the well bore. The collar was designed so that when positioned in the well essentially all upward flow originating below that point passed through the cylinder. Rotation of the meter vanes was signaled at the surface by clicks in a pair of headphones, the frequency of the clicks being proportional to the rate of flow. By positioning the meter at different depths in the well while the well was being pumped with a shallow-well suction pump, it was possible to determine fairly accurately the depth of the contributing zones. Wells 1, 3, 4, 5, and 6 (see fig. 2) were tested at rates ranging from 47 to 57 gal/min (3 to 4 l/s). Practically all detectable flow originated between the depths of 90 and 140 ft (27 and 43 m). Allowing for a margin of error in the testing technique, it is estimated that the principal contributing zone is between the depths of 80 and 150 ft (24 and 46 m). No flow was detected from a depth greater than 150 ft (46 m), that is,

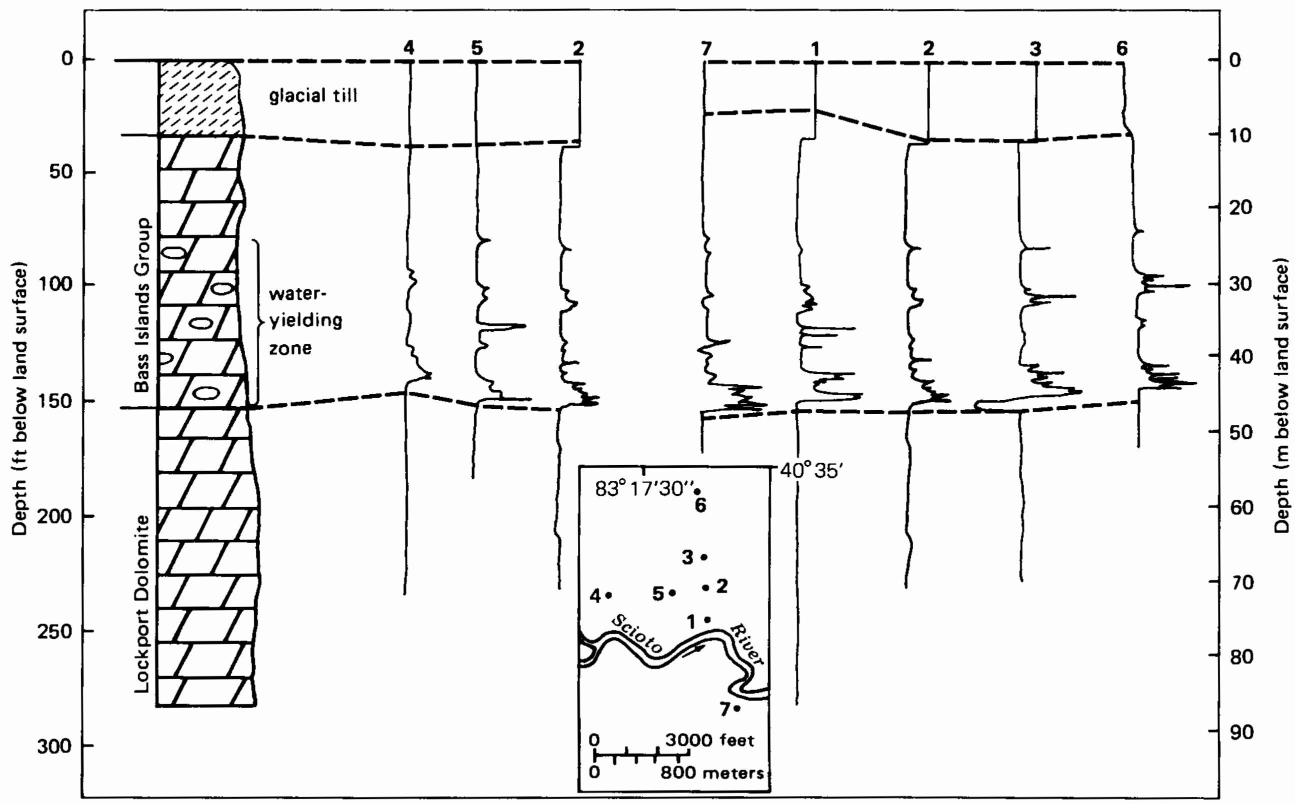

FIGURE 2.—Generalized geologic section at the test site and caliper logs of test wells showing position of principal water-yielding zone.

from the Lockport Dolomite.

The carbonate rocks at the test site are overlain by glacial till ranging in thickness from 22 to 37 ft (6 to 11 m), averaging 31 ft (9 m). The till is claylike, uncommonly dense, and contains few stones and stone fragments. The area is exceedingly flat and poorly drained. The soil is described on a soils map of the Ohio Division of Lands and Soil (Ritchie, 1973) as derived from "high lime glacial lake sediments." According to Ohio Division of Wildlife officials, the land is within a former wetland prairie, one of the larger prairies in Ohio at the time of settlement.

The ground dries out in summer and develops cracks up to an inch or more wide and at least several inches deep. However, the cracks do not seem to facilitate movement of water between the surface and the underlying carbonate rocks because the cracks evidently become closed a short distance beneath the surface. Prior to state acquisition of the land in 1958, farming was marginal because of the poor drainage. Some field tiles were laid as deep as 6 ft (2 m) in efforts to drain the land.

OCCURRENCE OF GROUND WATER

Ground water in the carbonate rocks is under artesian pressure because the water is confined by glacial till. The carbonate rocks and the regional slope rise to the southwest. The head in the aquifer is derived from this generally higher terrain, which includes the highest point in the state (1,549 ft (472 m) in altitude) located 25 mi (41 km) southwest of the test site. As shown by the potentiometric contours in figure 3, ground water moves from southwest to northeast in the site area. The hydraulic gradient was about 3 ft/mi (1 m/km) in the vicinity of the test wells in September 1973.

Ground-water levels at the test site range from 3 to 10

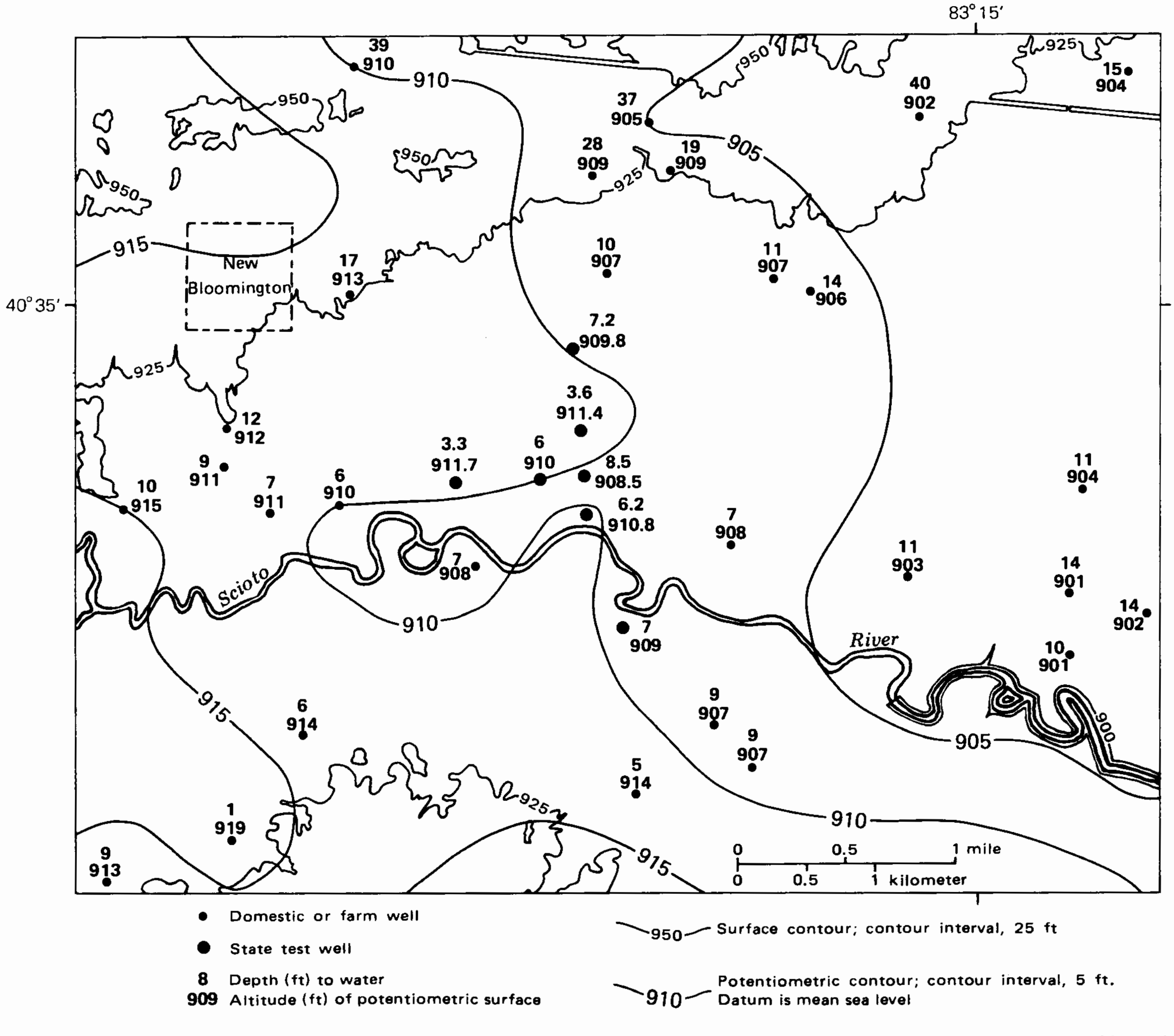

FIGURE 3.—Map of the Big Island Wildlife Area showing topography and contours on the regional potentiometric surface, September 20, 1973. Topography generalized from U.S. Geological Survey New Bloomington and Marion West 7½-minute quadrangle maps.

ft (1 to 3 m) below the surface depending on the season. The test site is considered to be in a discharge area because water levels in the deeper wells, tapping the more productive part of the aquifer, are higher than water levels in shallower wells open in the upper part of the carbonate rocks. The condition indicates general upward flow. Prior to each of the aquifer tests the water levels in the deeper wells ranged from a fraction of a foot to more than 2 ft (1 m) above the levels in the shallower wells.

Water levels in the till are generally below those in the carbonate rocks, indicating discharge from the aquifer through the overlying till either by outflow to surface streams or by evapotranspiration. Locally, water levels in the till may be higher than levels in the carbonate rocks, indicating inflow to the aquifer. This condition was observed at the test site at the time of the tests. The cause is believed to have been the artificial flooding, which produced a locally high water table in the till.

CHARACTERISTICS OF THE TEST WELLS

The first well drilled at the Big Island site was well 1, originally designated CPBR-4, and was completed in January 1971 as part of the state central Ohio water-plan study. It was anticipated that the well might be used to augment pumpage from the Scioto River; however, plans had not yet been made for large-scale aquifer testing using multiple wells.

Well 1 was drilled to the depth of 290 ft (88 m) and ultimately became the deepest well at the Big Island site. The entire carbonate-rock section beneath 22 ft (7 m) of glacial till was penetrated, and the well bottomed 5 ft (2 m) into the underlying shale. The field geologist reported that about 70 percent of the water produced during the drilling entered the well just above 150 ft (46 m). This permeable section is the Newburg zone. At the site of well 1, the base of the Newburg zone is 3 ft (1 m) above the top of the Lockport Dolomite (see fig. 2).

Other wells in the test pattern were drilled in the spring and summer of 1972. Well depths and other data are given in table 1; the layout of the test pattern is shown in figure 1.

Wells 1, 2, and 3 were constructed with a casing 12 inches (300 mm) in diameter from the surface to bedrock; holes 10 inches (250 mm) in diameter were continued into the carbonate section. Well 2 was 234 ft (71 m) deep and later was enlarged by reaming to 12 inches (300 mm) in diameter to the depth of about 165 ft (50 m).

Fully penetrating observation wells, 6 inches (150 mm) in diameter, were drilled at sites 4, 5, 6, and 7. In addition, shallow wells open in the top of the carbonate rocks were drilled at sites 4, 5, and 6. The shallow wells, designated by the letter A following the number, are located 5 to 10 ft (2 to 3 m) from the deep wells.

Wells 1, 2, and 3 are spaced 1,000 ft (305 m) apart on an approximate north-south line. With well 2 as the center, wells 1, 3, and 5 are equally spaced on a semicircle with a 1,000-ft (305-m) radius. Wells 4 and 6, to the west and north, respectively, are 3,000 ft (914 m) from well 2. Well 7 is 4,000 ft (1,219 m) south of well 2 and slightly east of the line that passes through wells 1, 2, 3, and 6 (see fig. 1). Spacing at less regular intervals would have been preferable as it would have resulted in better distribution of the datum points on distance-drawdown graphs.

TABLE 1.—*Characteristics of wells drilled at the Big Island test site*

Well no.	Diameter (in)	Altitude of land surface (ft)	Depth (ft)	Cavity zone *from to* (ft)	Depth to bedrock (ft)	Depth to Lockport Dolomite (ft)
1	10	916	290	82-149	22	153
2	10[1]	917	234	82-151	37	153
3	10	915	230	82-148	34	152
4	6	915	236	92-140	36	147
4A	6	915	58		36	
5	6	916	170	79-148	36	152
5A	6	916	57		36	
6	6	917	170	77-146	27	150
6A	6	916	52		27	
7	6	916	172	77-153	24	156

[1]Well 2 was enlarged to 12 inches in diameter to the depth of 165 ft.

PRELIMINARY AQUIFER TESTING

The 90-day aquifer test using well 2 began October 17, 1973, and provided most of the data for this report. However, shorter term tests also were made at the Big Island site.

In the spring and summer of 1972, step-drawdown and 1-day constant-rate tests were made by pumping wells 2 and 3. The most useful test was that of well 3. Profiles of the cone of depression were determined from measurements made in two nonpumping wells along lines nearly at right angles to each other. The pumping rate was 599 gal/min (38 l/s).

The 90-day test was the second attempt to make a long-term constant-rate test using well 2. Pumping was started on September 24, 1973, but power to the pump failed after 9 days. The pump was restarted October 17 after ground-water levels had been given time to recover. During the aborted 9-day test, which was at a higher pumping rate—494 versus 369 gal/min (31 versus 23 l/s)—drawdown measurements were made in all observation wells at short time intervals in the early part of the test. Datum plots from these measurements gave good definition of the early time-drawdown relationship. Consequently, fewer "rapid" measurements, those taken in the first few minutes and hours of testing, were made in the early part of the 90-day test.

EFFECTS OF PUMPING PAIRED SHALLOW AND DEEP WELLS

Pumping the paired shallow and deep observation wells gave good evidence of the contrast between the poorly conductive upper part of the carbonate-rock aquifer and the highly conductive lower part. Well 4, a fully penetrating well, was pumped at the rate of 47.5 gal/min (3 l/s) for 1 hour with a drawdown of 12.3 ft (4 m). The drawdown in well 4A, located about 10 ft (3 m) from well 4 and open only to the top 20 ft (6 m) of the aquifer, was 0.25 ft (0.1 m). One-hour tests of fully penetrating wells 5 and 6 at comparable pumping rates caused no discernible lowering of water levels in the shallow wells at those sites.

Despite the locally poor connection between the upper and lower parts of the carbonate-rock aquifer, the water level in shallow wells 5A and 6A responded to the pumping of well 2 very much as did the fully penetrating wells at those sites. However, drawdown was not as great in the shallow wells as in the deep wells.

At site 4, however, drawdown was greater in the shallow well than in the fully penetrating well. The initial response to pumping was observed in both well 4 and well 4A in about 10 minutes. These wells are 3,000 ft (914 m) from the pumped well. The observed initial-response time at site 6, also 3,000 ft (914 m) from the pumped well, was about 12 minutes in the fully penetrating well and about 60 minutes in the shallow well. At site 5, 1,000 ft (305 m) from the pumped well, the fully penetrating well responded to pumping within 5 or 10 seconds; the shallow well took a few tens of seconds longer to respond.

The fact that drawdown was greater in the shallow well at site 4 and in the deep well at site 5, located directly between the pumped well and site 4, is evidence of the heterogeneity of the aquifer, a condition verified by other evidence from the aquifer tests.

SPECIFIC-CAPACITY TESTS

A method of determining changes in aquifer properties is the comparison of specific-capacity data. The specific capacity of a well is its yield per unit of drawdown, commonly expressed as gallons per minute per foot of drawdown. Because of well losses, which increase disproportionately with higher pumping rates, comparison of specific capacities should be made for comparable pumping rates and periods of pumping. Individual tests of all but one of the test wells were made at pumping rates that ranged from 47.5 to 57 gal/min (3 to 4 l/s). The drawdown in each instance was determined after 1 hour of pumping.

Results of specific-capacity tests (table 2) indicate that the aquifer is most permeable in the vicinity of well 2. Permeability also is relatively high in the vicinity of wells 1 and 3. The permeability is relatively low in the vicinity of wells 4 and 6. The specific capacity of well 4 was significantly lower than that of the other wells.

TABLE 2.—*Specific capacities of the test wells*

Well no.	Pumping rate (gal/min)	Specific capacity (gal/min/ft)
1	57	51.8[1]
2	369[2]	85.8
3	55	55.0
4	47.5	3.9
5	53	40.1
6	54	22.2

[1] Specific capacity was 26 gal/min/ft at pumping rate of 500 gal/min.
[2] Not pumped at lower rate.

DYE-TRACER EXPERIMENT

On the first day of the aborted 9-day aquifer test and on the first day of the 90-day test 23 days later, a soluble red dye, Rhodamine WT, was injected into well 5, located 1,000 ft (305 m) from the pumped well. The dye was poured through 100 ft (30 m) of tubing so that it entered the well opposite the cavity zone (see table 1). The injection of the dye, 1 gal (4 l) of concentrated solution, was followed by the addition of 50 gal (189 l) of water, which forced the dye solution out of the well bore into the aquifer.

The purpose of the experiment was to see whether the dye would move through the aquifer to the pumped well, and if so, how long it would take. The experiment was performed even though preliminary calculations relating the pumping rate to the estimated volume of water in the aquifer indicated the dye was unlikely to appear in the pumped well during the test. It was important to learn whether water might move in preferred flow paths at much faster rates than indicated by the calculations. A fluorometer to detect the dye was set up at the pumped well (well 2). The effluent from the well was tested almost daily for much of the 9- and 90-day testing periods, and no dye was detected.

Calculations of ground-water velocity based on the observed hydraulic gradient indicated that by the end of the 90-day test the dye masses would have moved no more than about 35 ft (11 m) from well 5, the first mass leading the second by about 5 ft (2 m). Subsequent evidence confirmed that the dye did not move far from the well into which it was injected. This evidence was obtained in late May 1974, approximately 18 weeks after the end of the 90-day test, when well 5 was pumped at a low rate in connection with current-meter flow tests. The well was pumped at the rate of 53 gal/min (3 l/s) for 1 hour. The pumped water appeared turbid for the first minute or two of pumping but soon cleared. After 3 or 4 additional minutes the water began to redden, obviously because of the second dye, and within a few minutes it became dark red in color. After 5 or 10 minutes the color faded and the water became almost clear again. Then, 20 minutes from the start of pumping, the water again became dark red, leaving little doubt of the appearance of the first dye dose, the one that had been injected 23 days prior to the second dose. The water began to clear in another few minutes, and by the end of the hour-long test the color was almost gone.

Both dye masses were brought to the surface by pumping about 3,000 gal (11,400 l) of water, indicating that the dye had remained in the immediate vicinity of well 5 throughout the test and for many weeks thereafter. On the basis of uniform radial flow, failure of the dye to move as far as anticipated is further evidence of aquifer heterogeneity.

1972 TEST OF WELL 3

In analytical techniques involving graphs of drawdown versus distance it is desirable to plot the drawdown in the pumped well. However, the observed drawdown must be adjusted to eliminate the well-loss component. Well loss, that part of the drawdown due to turbulent flow, is commonly determined from a step-drawdown test in which a well is pumped for short periods at successively higher rates. Turbulent flow, which occurs in the well bore and sometimes in the aquifer close to the well bore, causes a disproportionately larger drawdown for each higher pumping rate. The remainder of the drawdown, that due to laminar flow, is directly proportional to the pumping rate and depends only on aquifer characteristics and time of pumping.

A step-drawdown test of well 3 utilized four 1-hour pumping periods at rates of 446, 599, 752, and 901 gal/min (28, 38, 47, and 57 l/s), respectively. Data were analyzed by

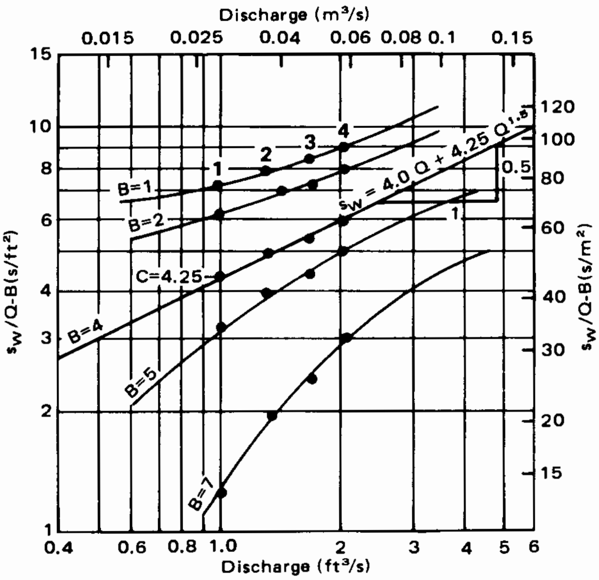

FIGURE 4.–Graphical solution of step-drawdown test of well 3 (s_w = BQ + CQ^n). See text for explanation of equation and symbols. All values on graph (B = 1, etc.) in English units.

the method devised by Rorabaugh (1953) and illustrated in figure 4. The drawdown components in a pumped well are expressed as:

$$s_w = BQ + CQ^n$$

where

s_w = drawdown in the pumped well,
B = formation factor,
C = "well loss" constant,
Q = discharge of the pumped well,

and

n = exponent for turbulent flow.

The exponent for turbulent flow (n) is assumed to be 2 in some analytical methods (Jacob, 1947; Bruin and Hudson, 1955); however, Rorabaugh (1953) shows that the value of this exponent is variable. In his method, step discharge, in ft³/s, versus the quantity (s_w/Q) - B, where B is selected by trial and error to find the value which produces a straight-line plot, are plotted on logarithmic coordinates. As shown in figure 4 the value B = 4 resulted in a straight-line plot of the data from well 3; other values for B produced curved lines, indicating incorrect solutions to the drawdown equation. The slope of the straight-line plot equals n - 1, therefore n equals 1.5; C is obtained from the intercept of the straight line with the vertical line where Q = 1.

The step-test analysis was used to determine the magnitude of the drawdown due only to aquifer characteristics in well 3 at the end of the constant-rate test. This adjusted drawdown value and drawdowns recorded in the observation wells are plotted on a semilogarithmic graph in figure 5 to produce a profile of the cone of depression after 1 day of pumping at the rate of 599 gal/min (38 l/s). Data from wells 1 and 2 define the profile along a line extending

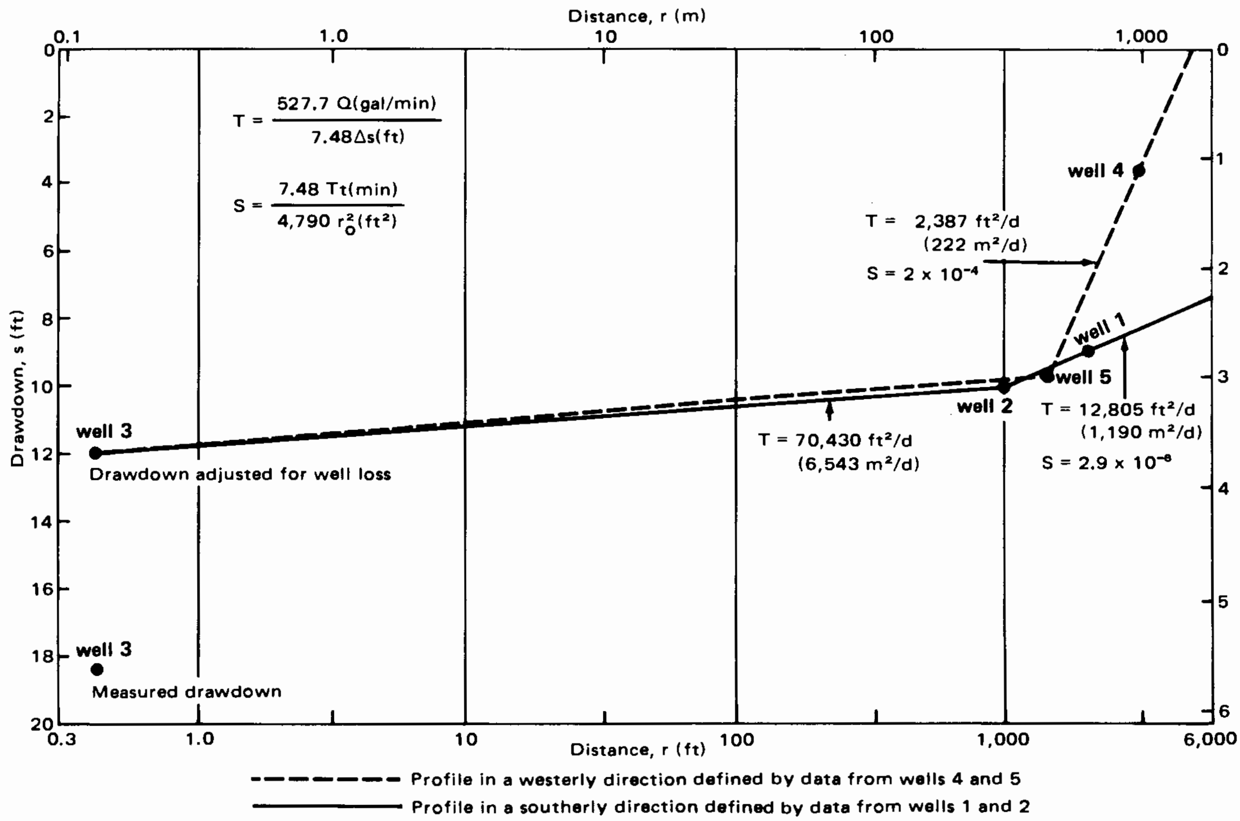

FIGURE 5.–Semilogarithmic plot of drawdown versus distance after one day of pumping from well 3. See text for explanation of equations and symbols.

south from the pumped well. Although wells 4 and 5 are not strictly on line with the pumped well, data from these wells define the cone in a westerly direction from the pumped well. The most significant feature of the graph (fig. 5) is the flatness of the cone of depression in the vicinity of wells 1, 2, 3, and 5. Values for transmissivity and the storage coefficient were calculated from the straight-line graph segments using a variant of the Theis nonequilibrium formula.

The nonequilibrium formula introduced by Theis (1935) is described in publications by Wenzel (1942, p. 88-89), Ferris and others (1962, p. 92-98), Lohman (1972, p. 15-19), and in other publications of the U.S. Geological Survey. A succinct description of the formula and examples of its use are given by Lang (1960, p. 356-364). The Theis nonequilibrium formula is:

$$s = \frac{114.6Q}{7.48T} W(u)$$

where

s = drawdown (ft) at any point,
Q = discharge rate (gal/min) of pumped well,
T = transmissivity (ft²/d) of the aquifer,

$$W(u) = -0.5772 - \log_e u + u - \frac{u^2}{2 \cdot 2!} + \frac{u^3}{3 \cdot 3!} - \frac{u^4}{4 \cdot 4!} \cdots,$$

and

$$u = \frac{2{,}693r^2S}{7.48Tt},$$ in which S is the storage coefficient of the aquifer; r is the distance, in feet, from the pumping well to a point at which drawdown, s, is determined; and t is the pumping time, in minutes.

Transmissivity, preferred over the term coefficient of transmissibility, is a measure of the property of the aquifer to transmit water. As defined by Lohman (1972, p. 6) transmissivity is "the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient." The storage coefficient, defined originally by Theis (1935), is given by Lohman (1972, p. 8) as "the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head." The storage coefficient is a dimensionless number that is expressed also as a percentage. Lohman (1972, p. 8) states, "the storage coefficient of most confined aquifers ranges from about 10^{-5} to 10^{-3} and is about 10^{-6} per foot of thickness. In contrast the specific yield [storage coefficient] of most unconfined aquifers ranges from about 0.1 to about 0.3 and averages about 0.2."

Cooper and Jacob (1946) recognized that when u becomes small (less than about 0.01) the sum of the terms in $W(u)$ beyond $\log_e u$ becomes insignificant; u becomes small when time, t, becomes large or when distance, r, is small. In the region where u is small, semilogarithmic plots of drawdown, s, on an arithmetic scale versus either distance, r, or time, t, on a logarithmic scale will yield a straight-line graph. The slope of the straight line is used to calculate the transmissivity, T, and the intercept of the straight line with either zero distance or zero drawdown, depending on the graph, is used to calculate the storage coefficient. The equations used with distance-drawdown and time-drawdown graphs, respectively, are:

$$T = \frac{527.7Q}{7.48\Delta s}$$

$$S = \frac{7.48Tt}{4{,}790r_o^2},$$

and

$$T = \frac{264Q}{7.48\Delta s}$$

$$S = \frac{7.48Tt_o}{4{,}790r^2},$$

where

T = transmissivity (ft²/d),
S = storage coefficient,
Q = discharge rate (gal/min) of well,
r = distance (ft) from pumped well to observation well,
r_o = intersection (ft) of straight-line slope with zero-drawdown axis,
t = time (min) since pumping began,
t_o = intersection (ft) of straight-line slope with zero-drawdown axis,

and

Δs = change (ft) in drawdown over one log cycle.

Among the more important assumptions in the use of the nonequilibrium formula are that the aquifer is homogeneous and isotropic and that the water removed from storage is discharged instantaneously with a decline in head.

Use of the straight-line distance-drawdown method assumes that the ground-water gradient has stabilized between points on the cone at which the drawdown was determined. It is unlikely that this condition was met during a pumping period of 1 day. The slopes of the straight-line plots in figure 5 are very flat between wells 2 and 3 and between wells 3 and 5, indicating relatively high values of transmissivity in the region close to the pumped well. The slopes are comparatively steep between wells 1 and 2 and between wells 4 and 5, indicating relatively low transmissivity at distances farther from the pumped well.

The marked changes in slope of the cone are the result of aquifer heterogeneity. Heterogeneity is reflected also in the widely disparate and somewhat unrealistic values determined for the storage coefficient.

The storage coefficient calculated from the drawdown graph (fig. 5) between wells 1 and 2 is about 3×10^{-8}, and the coefficient based on the drawdown graph between wells 4 and 5 is about 2×10^{-4}. The first of these values is too small to fit the range of values (10^{-5} to 10^{-3}) given by Lohman for artesian aquifers; the second falls within this range. The storage coefficient, were it calculated from either the segment of the semilogarithmic graph between wells 2 and 3 or the segment of the graph between wells 3 and 5 (fig. 5), would be several orders of magnitude less than the lower of the two values given above.

The distance-drawdown method using wells located within the region of high transmissivity does not give a reasonable value for the storage coefficient, suggesting that this region is primarily a zone of transmission with little release of water from storage. Analysis of the distance-drawdown slope beyond this region in the direction of well 4 yields a reasonable value for the storage coefficient, suggesting that most storage release takes place beyond the region of high transmissivity.

The distance-drawdown analysis of the 1-day aquifer test using well 3 can be summed up as follows: (1) the aquifer is heterogeneous; (2) a zone of relatively high hydraulic conductivity exists in the vicinity of wells 1, 2, 3, and 5; and (3) most of the water released from storage in the aquifer comes from the area beyond the region of high hydraulic conductivity.

FIELD FACILITIES AND INSTRUMENTATION

After the preliminary testing, preparations were made for a long-term aquifer test using well 2. A special power line was extended to the well site, a pump house was built, and an electrically powered turbine pump capable of pumping water at an estimated 1,300 gal/min (82 l/s) was installed in the well. A propeller-type flowmeter was installed in the discharge line to determine the pumping rate and quantity of water pumped. The pumping rate was recorded continuously on a clock-driven circular chart; however, more accurate determinations of the pumping rate during the subsequent tests were made by relating differences in individual meter readings to selected time periods.

The water level in the pumped well was measured by a chalked steel tape or electric sounder which could be lowered into the casing through a drop pipe that was 0.75 inch (20 mm) in diameter and extended from the pump base to a point just above the impellers.

Water-level recorders were installed on all state-owned wells and operated throughout the testing period. The recorders were equipped with 1:10 float gears, meaning that a 10-ft (3-m) change in water level produced a 1-ft (0.3-m) change on the recorder chart. The recorders were operated with daily gears for the first 2 or 3 days of the 9-day test, after which they were switched to weekly gears.

Drawdown measurements made in the first minutes and hours of the tests were timed with stopwatches and recorded manually on the charts. Figures 6, 7, 8, and 9 are representative hydrographs showing fluctuation of water levels during the 9-day and 90-day tests.

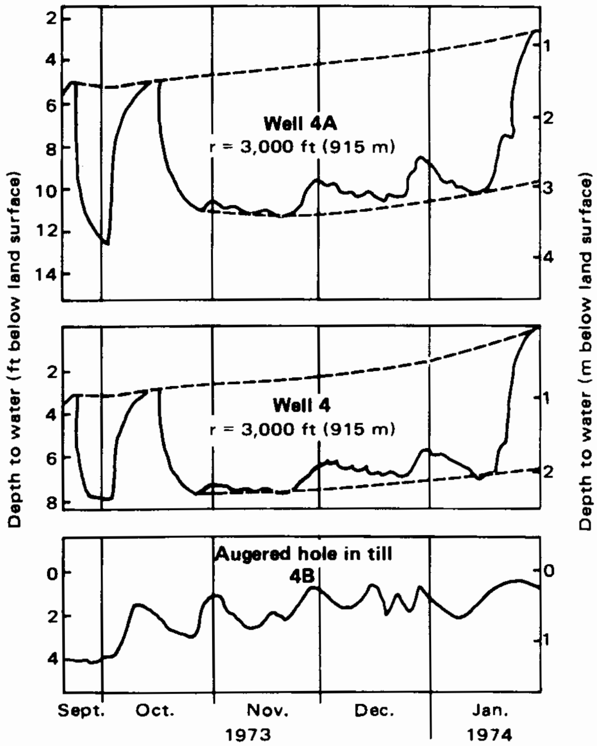

FIGURE 7.—Hydrographs of wells 4 and 4A and of augered hole in till showing fluctuation of water levels during aquifer tests; *r*, distance from pumped well 2.

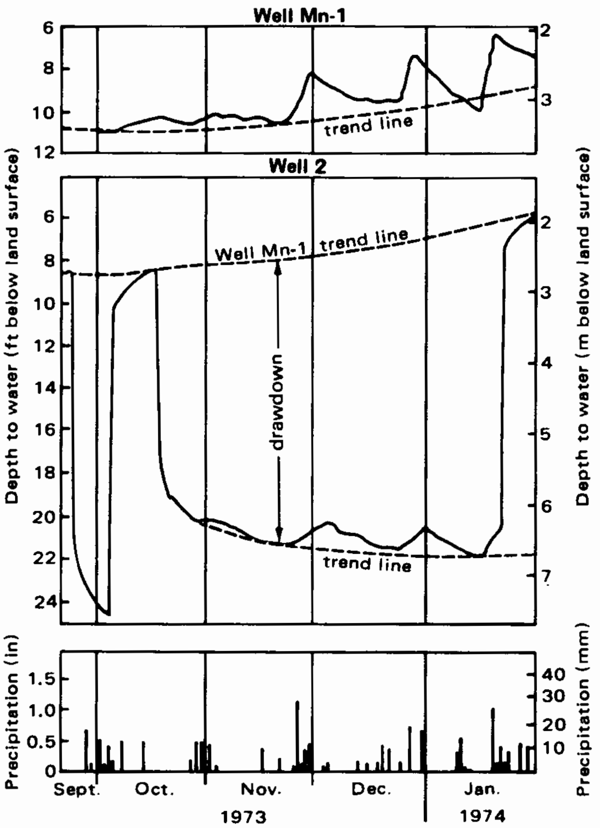

FIGURE 6.—Hydrographs of state observation well Mn-1 and pumped well 2, showing fluctuation of water levels during aquifer tests. Also shown is precipitation at LaRue.

Periodically during both tests, water-level measurements were made in private wells located on the periphery of the wildlife area. This was done to determine drawdown and to determine whether water levels were being lowered to the point that local residents using shallow-well suction pumps might have difficulty obtaining water. It was known from preliminary testing that pumping effects would extend over a large area. The pumping rates, 494 gal/min (31 l/s) for the first test and 369 gal/min (23 l/s) for the second test, were kept low to avoid excessive lowering of water levels in nearby farm wells.

During the 9-day and 90-day tests a water-level recorder was operated continuously on state observation well Mn-1, which taps the carbonate-rock aquifer at the village of LaRue about 5 mi (8 km) west of the test site. Because it was far enough away to be unaffected by the test pumping, the hydrograph of the Mn-1 well (fig. 6) was similar to the prepumping hydrographs of the test wells and was used as a base for determining drawdown in the test wells due to pumping.

The water table in the glacial till was monitored during the tests to determine whether it was affected by pumping. Piezometer holes, 8 to 9 ft (2 to 3 m) deep, were hand augered into the till at each of the test-well sites and kept open with short lengths of plastic pipe. Five holes, three of which were located in a shallow ditch inside the diked area,

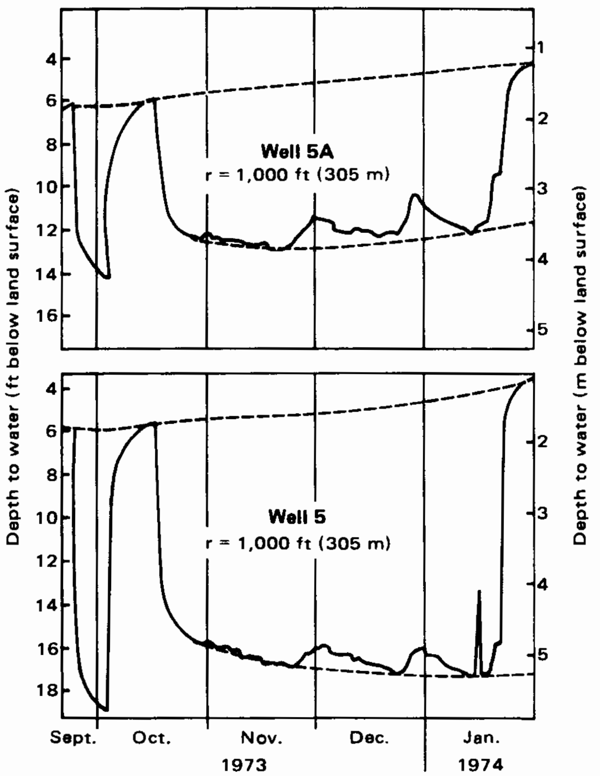

FIGURE 8. Hydrographs of wells 5 and 5A showing fluctuation of water levels during aquifer tests; *r*, distance from pumped well 2.

were augered in the immediate vicinity of the pumped well (fig. 9). The ditch was partly filled with water which rose during the tests and eventually surrounded the shallow piezometer tubes. The tubes remained water tight, however, and there was a measurable difference between the water level inside the tubes and the level on the outside throughout the test. The water level in these three holes (2B, 2C, 2D) fluctuated very little during pumping (see fig. 9). In contrast, the water level in the piezometer tubes outside the diked area fluctuated up to 5 ft (2 m) (see fig. 7). These were natural fluctuations caused by precipitation and did not appear to reflect pumping effects.

A recording barometer was operated at the test site. Fluctuations in atmospheric pressure corresponding to as much as 1 ft (0.3 m) or more of water, but usually much less than this, were recorded at various times over periods of 2 to 4 days (fig. 9). Barometric effects on the water levels were slight, and no corrections were made in computing drawdown.

Water pumped during the tests was discharged directly into the diked area. At the start of the tests in September 1973, the diked area was essentially dry except for water in the peripheral ditches from which material had been excavated to construct the dikes. By about the middle of October, precipitation, discharge from well 2, and pumpage from the Scioto River had produced fairly general flooding in the diked area. Water was standing in low places in other parts of the tract. As early as October 9 an inch or more of water covered a sizeable area in the vicinity of well 3. A few weeks later there were shallow pools of water in the vicinity of wells 5 and 5A. Once established, the pools were frozen and became snow covered after the first week of December. The pools persisted for the duration of the 90-day test, thus helping to maintain a locally high water table in the till.

9-DAY AND 90-DAY AQUIFER TESTS USING WELL 2

On September 24, 1973, at 1130 hours, pumping was started in well 2 at the rate of 494 gal/min (31 l/s) for what was intended to be a test of 30- to 90-day duration. A power failure during an electrical storm stopped the pump at approximately 0415 hours on October 4 after 9 days 16 hours 45 minutes of pumping. The pump was restarted at 1130 hours on October 17, 1973, after water levels had recovered, and continuous pumping was maintained for a little more than 90 days.

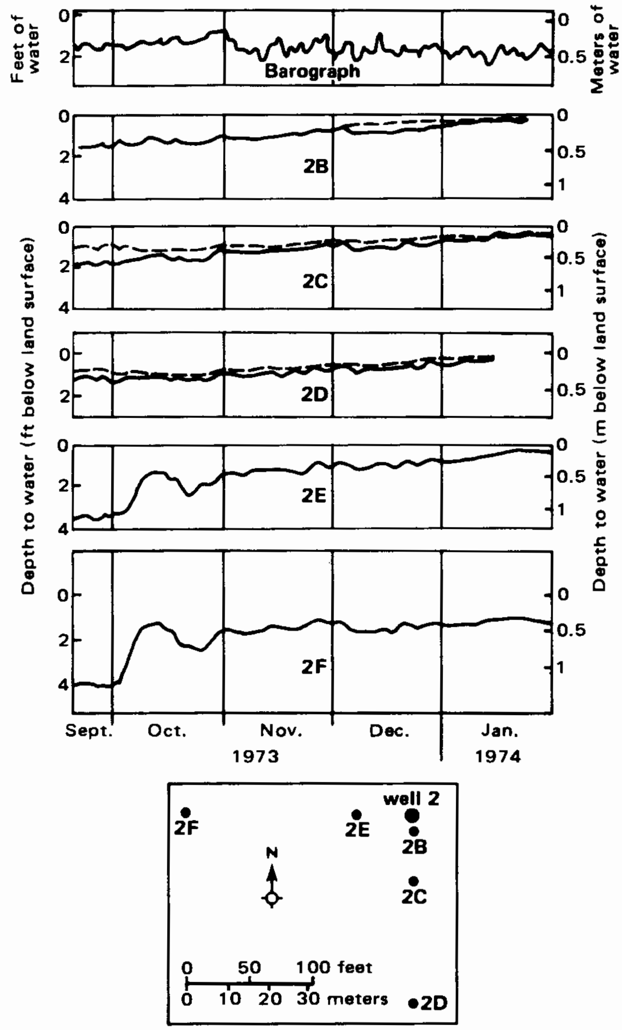

FIGURE 9.—Hydrographs of augered holes in till in the vicinity of well 2 showing fluctuation of water levels during aquifer tests. Dashed line indicates depth to water on outside of piezometer tube. Barograph is also shown.

TABLE 3.–*Drawdown in the test wells for selected pumping periods*

Time since pumping began (days)	Drawdown (ft)									
	Well 1	Well 2[1]	Well 3	Well 4	Well 4A	Well 5	Well 5A	Well 6	Well 6A	Well 7
	9-day test; Q[2] = 494 gal/min (31 l/s)									
2	11.4	14.2	11.0	4.6	5.6	11.5	6.2	6.4	3.6	5.9
3	11.5	14.8	11.2	4.8	6.2	11.7	7.1	7.2	4.3	6.7
4	11.8	15.2	11.5	4.9	6.6	12.0	7.5	7.5	4.7	7.0
5	12.0	15.4	11.7	4.9	6.9	12.2	7.7	7.8	4.9	7.2
7	12.3	15.7	12.1	4.9	7.1	12.6	8.0	8.2	5.2	7.6
9	12.5	15.9	12.3	4.9	7.2	13.0	8.2	8.4	5.4	7.7
	90-day test[3]; Q = 369 gal/min (23 l/s)									
2	7.9	9.9	7.7	3.4	4.4	8.2	4.8	4.8	2.4	4.3
3	8.5	10.4	8.2	3.9	4.9	8.8	5.3	5.4	3.0	4.9
4	8.9	10.9	8.7	4.2	5.2	9.1	5.8	5.8	3.4	5.3
5	9.3	11.1	8.9	4.4	5.5	9.3	6.0	6.1	3.7	5.5
7	9.5	11.5	9.3	4.7	5.8	9.7	6.3	6.4	4.0	5.9
9	9.9	11.8	9.6	4.9	6.0	10.2	6.6	6.8	4.2	6.2
10	10.1	11.9	9.7	5.0	6.1	10.3	6.7	6.9	4.3	6.3
20	10.7	12.5	10.4	5.1	6.4	10.9	7.0	7.5	4.7	6.8
25	11.0	12.8	10.6	5.2	6.5	11.2	7.2	7.7	4.9	6.9
30	11.2	13.1	10.9	5.2	6.6	11.4	7.3	7.9	5.1	7.1
42	11.7	13.7	11.3	5.2	6.8	11.8	7.5	8.2	5.1	7.4
68	12.3	14.4	12.0	5.5	7.0	12.5	7.6	8.4	5.1	7.6
90	12.9	15.2	12.6	5.9	7.0	13.1	7.7	8.5	5.1	7.7

[1] Pumped well.
[2] *Q*, pumping rate.
[3] Drawdown values after 30 days are less accurate than those for earlier times.

Little rain fell prior to and during the 9-day test, and antecedent ground-water levels were stable (see fig. 6). The time-drawdown data, especially data collected during the early part of the test when frequent measurements were made in all observation wells, are excellent.

The pumping rate for the 90-day test was 369 gal/min (23 l/s), about 25 percent lower than the rate for the 9-day test. Fewer "rapid" measurements were made in the observation wells during the 90-day test. Drawdowns were based, as in the 9-day test, on the control or background well Mn-1 located at LaRue (see fig. 6).

Table 3 lists drawdown values determined in the test wells for selected intervals during the aquifer tests. Rains in late October and early November produced minor fluctuations in ground-water levels but caused little difficulty in determining drawdowns accurately. Rains of greater intensity and duration occurring in the latter half of November and at various times in December and January caused larger fluctuations and made it more difficult to determine drawdowns with confidence. Values depend in large part on judgment used in applying the nonpumping data from well Mn-1 at LaRue. Drawdown values given for pumping time greater than about 30 days are much less accurate than the values given for earlier times. As will be shown later in describing the rate of growth of the cone of depression, the increase in drawdown in the latter two-thirds of the 90-day test may have been essentially zero.

DISTANCE-DRAWDOWN STRAIGHT-LINE ANALYSIS

Semilogarithmic plots of drawdown versus distance after 2 and 20 days of pumping (fig. 10) show that the shape of the cone of depression in the area near the pumped well became essentially stable after a relatively short period of pumping. The slope is relatively flat between the pumped well (well 2) and well 3, reflecting the region of high hydraulic conductivity around these wells noted previously in the test of well 3.

Noteworthy is the fact that values of transmissivity calculated from the flat portion of the graph between wells 2 and 3 in figure 10 are lower than the values based on the slope between the same two wells in figure 5. In one test, well 2 was pumped and in the other test, well 3 was pumped. The difference in transmissivity is probably due to the different shape and areal distribution of the region of high hydraulic conductivity with respect to each well.

The distance-drawdown curves (figs. 5, 10) suggest that the transmissivity is 47,000 to 70,000 ft^2/d (4,400 to 6,500 m^2/d) in the vicinity of wells 1, 2, 3, and 5, and 2,000 to 4,000 ft^2/d (190 to 370 m^2/d) farther from the center of pumping in the vicinity of wells 4 and 6.

SPECIAL TEST OF WELL 2

In an effort to learn more about the zone of high hydraulic conductivity near wells 1, 2, 3, and 5, a special test was made in which water-level measurements were taken in well 2 at short intervals for about 3 hours while it was being pumped at an average rate of 375 gal/min (24 l/s).

The drawdown data, corrected for well loss and adjusted to the average pumping rate, are plotted against time in figure 11. For comparison a segment of the Theis (1935) nonequilibrium type curve is shown matched to hypothetical plots of drawdown versus time for a well of

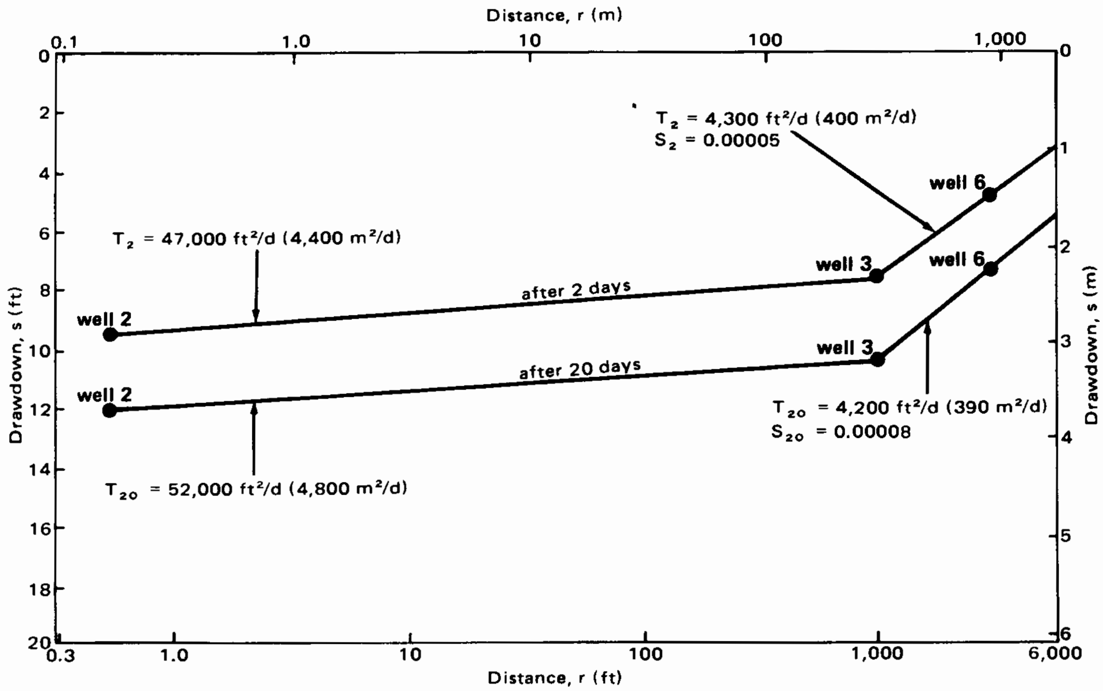

FIGURE 10.—Semilogarithmic graphs of drawdown versus distance after 2 days and 20 days of pumping from well 2. Drawdown in well 2 adjusted for well loss. See text for explanation of equations and symbols.

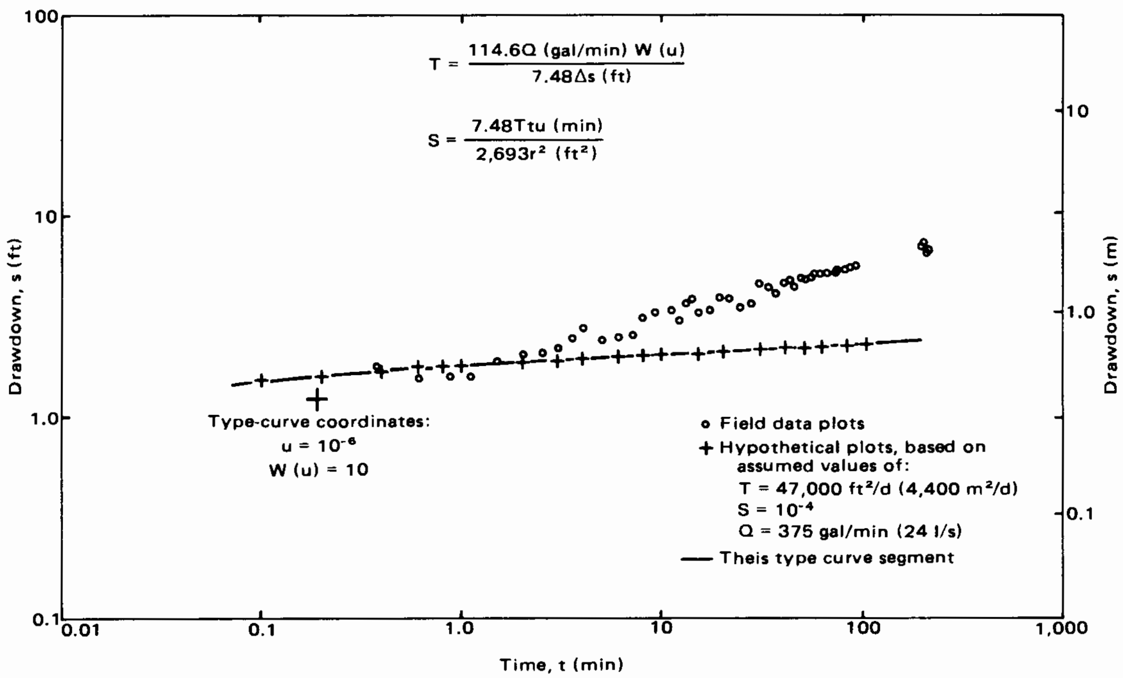

FIGURE 11.—Logarithmic graph of drawdown versus time in well 2. Drawdown adjusted for well loss. Also shown is a plot of hypothetical drawdown based on the Theis (1935) nonequilibrium formula. See text for explanation of equations and symbols.

6-inch (150-mm) radius being pumped at 375 gal/min (24 l/s) from an infinite aquifer whose transmissivity and storage coefficient equal 47,000 ft²/d (4,400 m²/d) and 10^{-4}, respectively. The transmissivity value is the same as the value obtained from the flat part of the distance-drawdown curve for a pumping period of 2 days in figure 10.

For such a small radius, time-drawdown data fall on the flat portion of the type curve after only a few seconds of pumping. Plots of the early-time field data also are essentially horizontal and match the hypothetical data plots for about the first 2 minutes. After the first 2 minutes the field data depart from the type curve in a manner somewhat analogous to the effect of a negative boundary. The early match of the field data with the type curve suggests that the aquifer constants determined from the close-in distance-drawdown data probably are valid for the region close to the pumped well.

A departure curve constructed from plots of the magnitude of the departures from the type curve (see Lang, 1960, p. 358) also matches a portion of the type curve. However, values of transmissivity and the storage coefficient determined from the departure curve differ by orders of magnitude from the selected values used in constructing the type-curve plots in figure 11. It is concluded, therefore, that these departures represent transition effects caused by the spread of the cone into less conductive parts of the aquifer.

The time-drawdown data from well 2 (fig. 11) indicate that the cone spreads rapidly beyond the region of high hydraulic conductivity, suggesting, as do the distance-drawdown data, that this region is primarily a zone of transmission. This highly transmissive zone is probably small compared with the aquifer as a whole.

ANALYSIS OF DRAWDOWN-CONTOUR MAP

Almost daily during the 9-day test, water-level measurements were made in farm and domestic wells in the test-site area. Figure 12 is a drawdown-contour map showing the

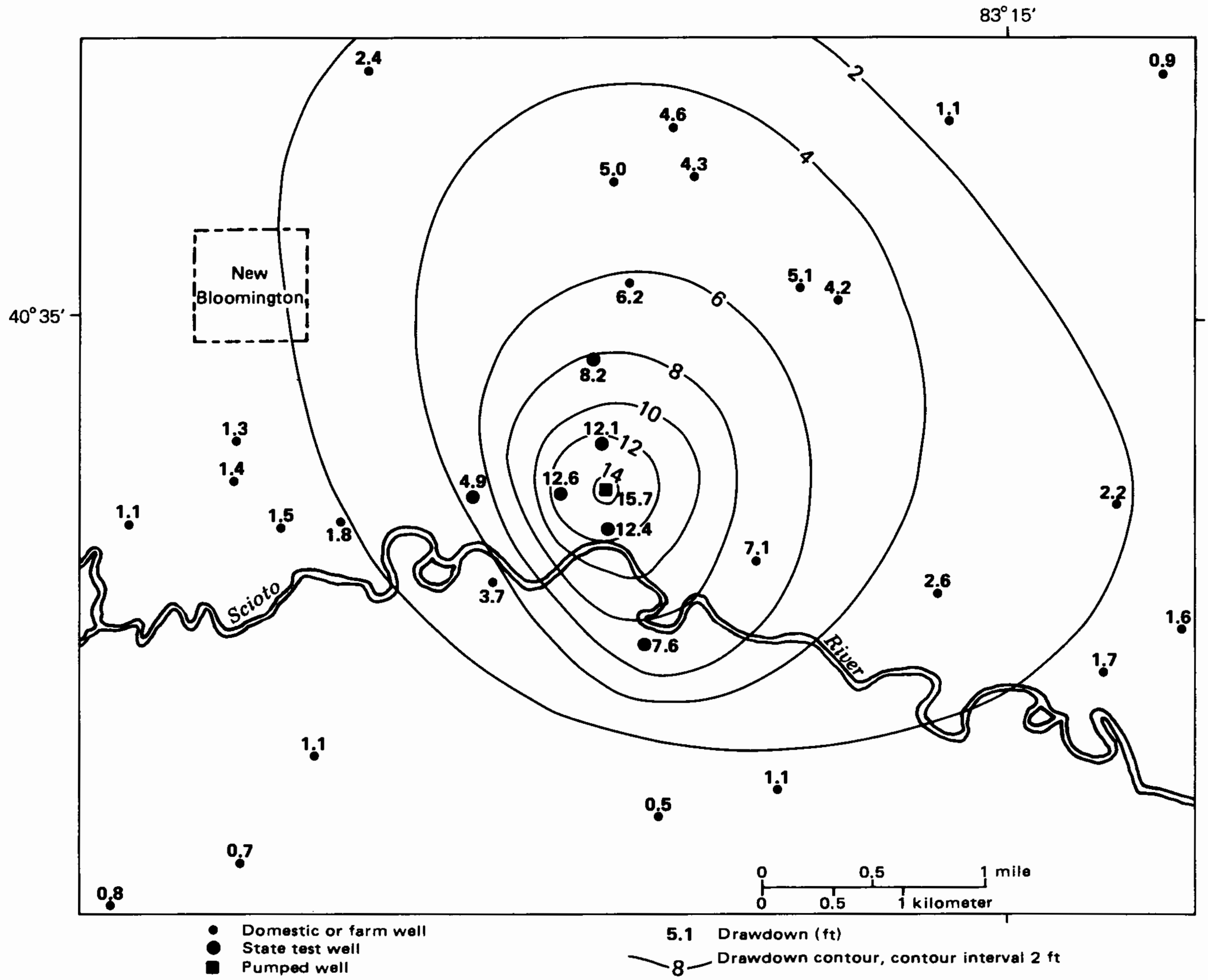

FIGURE 12.–Drawdown of the potentiometric surface on October 2, 1973, after well 2 had been pumped approximately 7 days at a rate of 494 gal/min (31 l/s).

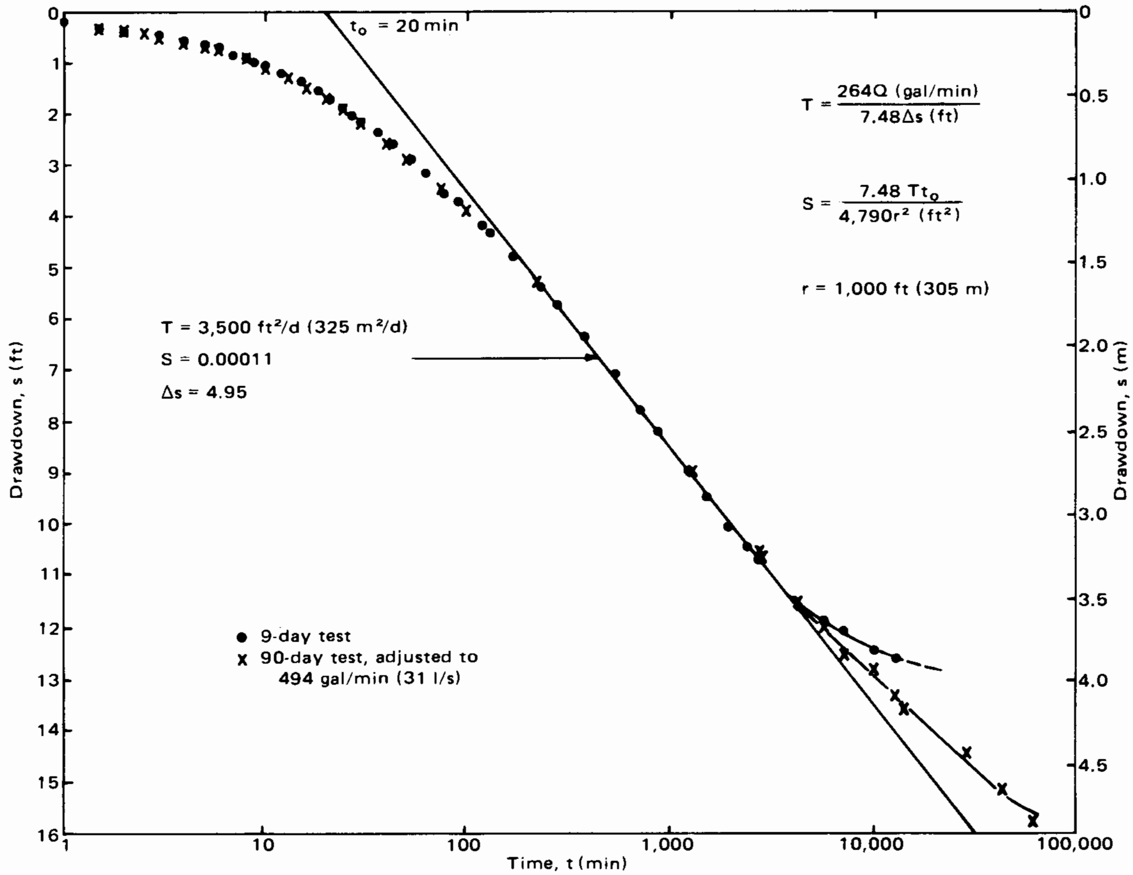

FIGURE 13.—Semilogarithmic graph of drawdown versus time in well 1 during 9-day and 90-day aquifer tests. See text for explanation of equations and symbols.

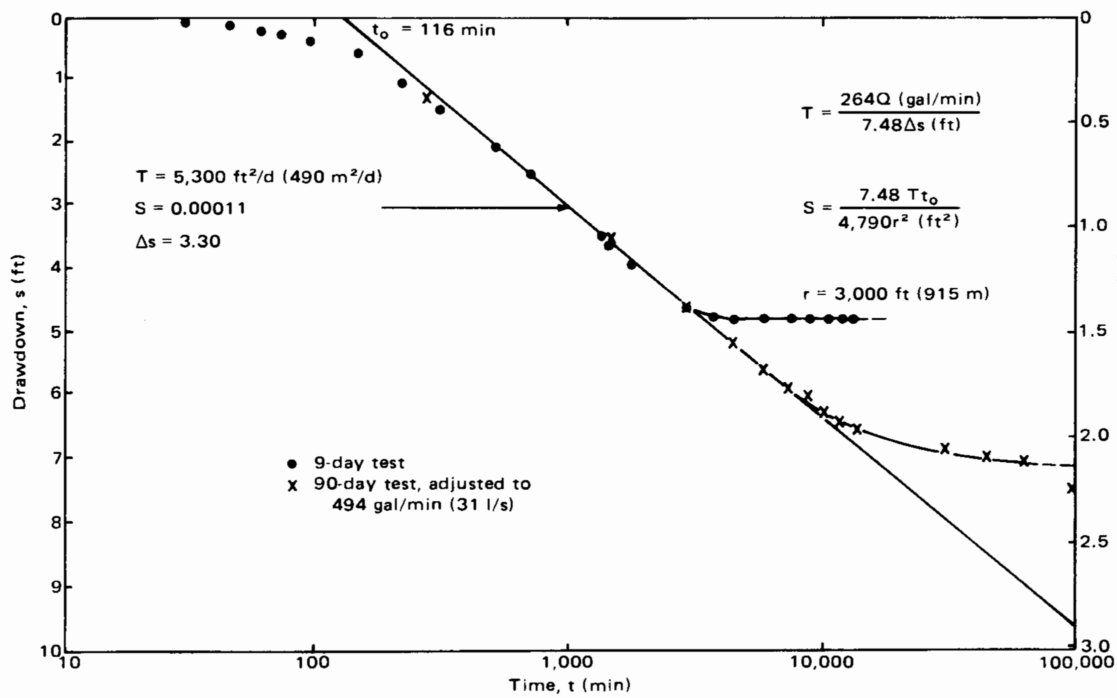

FIGURE 14.—Semilogarithmic graph of drawdown versus time in well 4 during 9-day and 90-day aquifer tests. See text for explanation of equations and symbols.

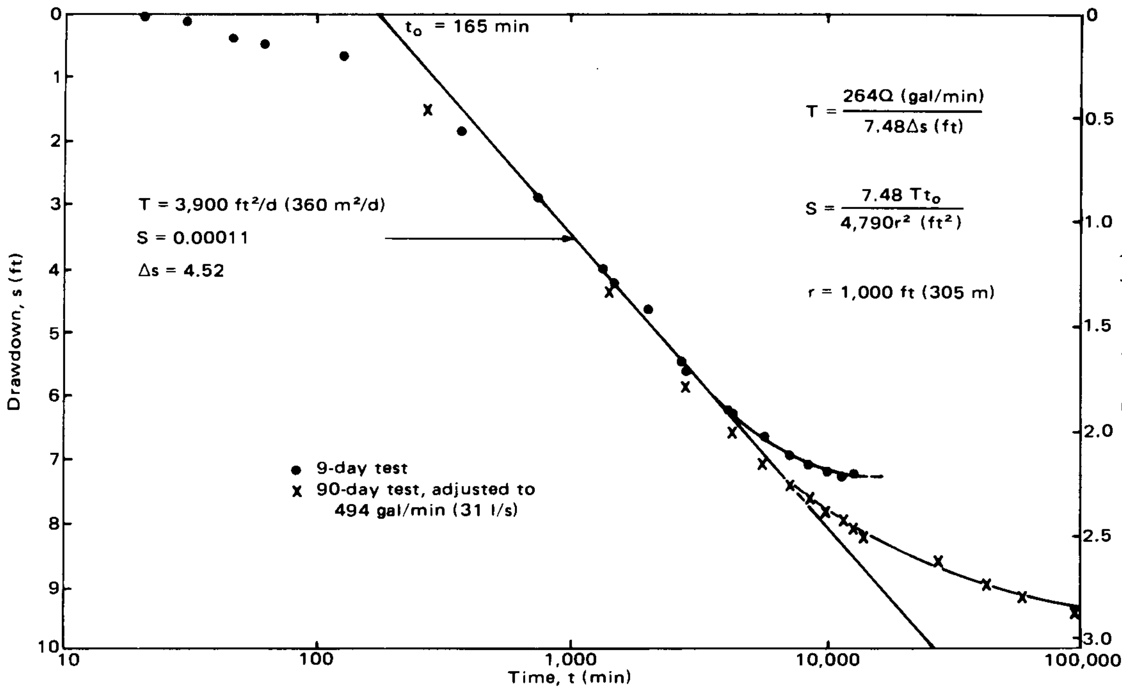

FIGURE 15.–Semilogarithmic graph of drawdown versus time in well 4A during 9-day and 90-day aquifer tests. See text for explanation of equations and symbols.

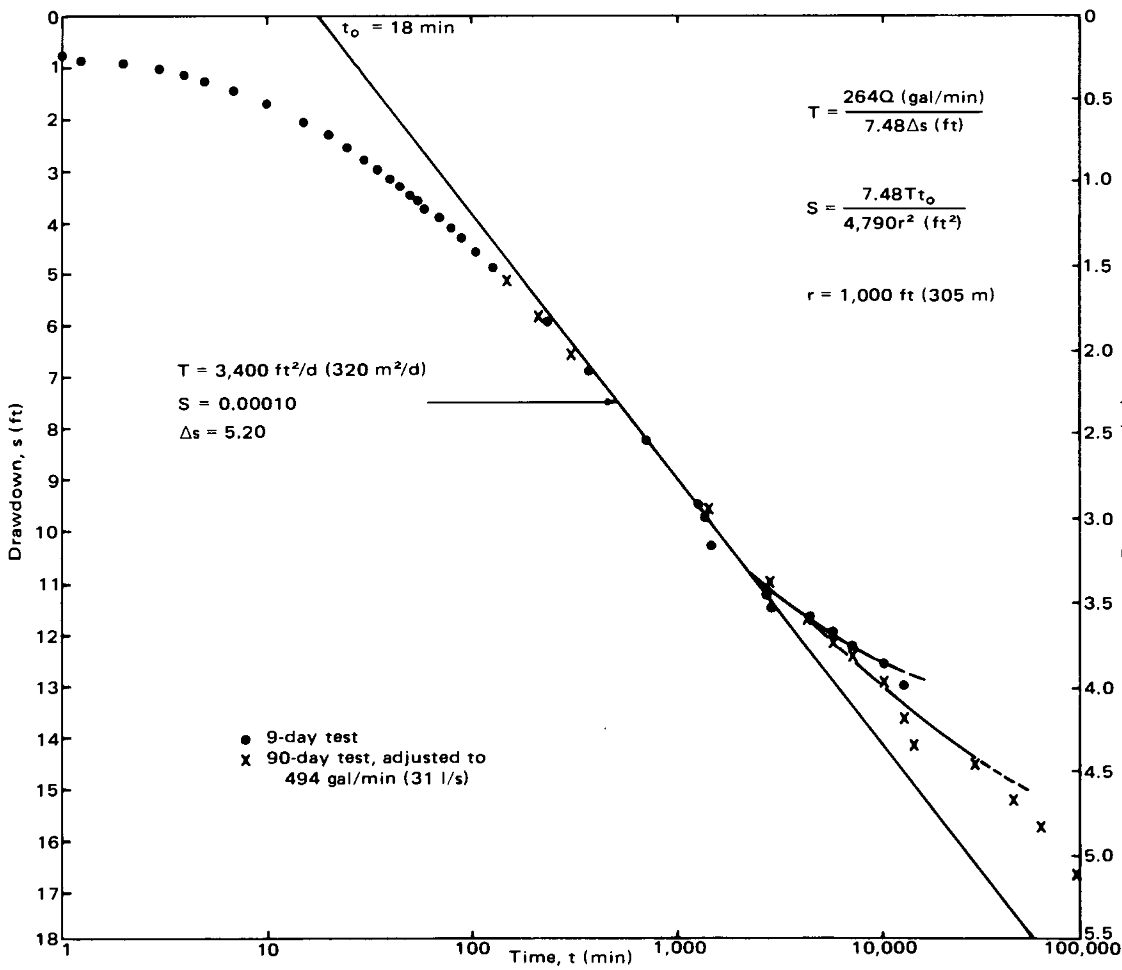

FIGURE 16.–Semilogarithmic graph of drawdown versus time in well 5 during 9-day and 90-day aquifer tests. See text for explanation of equations and symbols.

distribution of drawdown after approximately 7 days of pumping. The drawdown gradient is steepest on the southwest side of the pumped well in the area of relatively low hydraulic conductivity.

The drawdown map was used to determine the transmissivity in relatively distant parts of the cone of depression by using selected pairs of contours and the following relationship (Lohman, 1972, p. 47):

$$T = \frac{2Q\Delta r}{(L_1 + L_2)\Delta h}$$

where

T = transmissivity (ft^2/d),
Q = pumping rate (ft^3/d),
$L_{1,2}$ = length (ft) of closed contour,
$(L_1 + L_2)/2$ = average length (ft) of two adjacent contours,
Δh = contour interval (ft),

and

Δr = average distance (ft) between the two adjacent contours.

Using the 4- and 6-ft contours in the outer part of the cone, the results are:

$$T = \frac{2 \times 95{,}102 \text{ ft}^3/\text{d} \times 1{,}985 \text{ ft}}{(26{,}582 \text{ ft} + 40{,}965 \text{ ft}) \times 2 \text{ ft}} = 2{,}800 \text{ ft}^2/\text{d } (260 \text{ m}^2/\text{d}).$$

Although the transmissivity in the outer part of the cone is low, it may be fairly representative of the regional transmissivity of the aquifer.

TIME-DRAWDOWN STRAIGHT-LINE ANALYSIS

Figures 13, 14, 15, 16, and 17 are semilogarithmic graphs of drawdown versus time in wells 1, 4, 4A, 5, and 5A, respectively. These graphs represent the time-drawdown relationship at various points within the cone during the 9-day and 90-day tests. To better compare the data from the two tests, the drawdown values for the 90-day test were multiplied by the ratio of the discharges of the two tests (494/369 = 1.34) so that the sets of data could be plotted together.

On all graphs the transmissivity and storage coefficient were calculated from the straight-line portion beginning after 200 to 300 minutes and ending after 4,000 to 6,000 minutes, or between the time the plots began to fall on a straight line and the onset of recharge, which caused a change in the slope of the line. For the fully penetrating wells the average values for the transmissivity and storage coefficient are 3,600 ft^2/d (330 m^2/d) and 0.00011, respectively. Values for all wells are given in table 4.

TABLE 4.—*Transmissivity and storage coefficient determined from semilogarithmic time-drawdown graphs*

Well no.	Transmissivity (ft^2/d)	Storage coefficient (percent)
1	3,522	0.00011
2[1]	3,595	
3	3,385	0.00014
4	5,283	0.00011
4A	3,900	0.00011
5	3,353	0.00010
5A	3,790	0.00075
6	3,918	0.00010
6A	3,290	0.00037
7	3,632	0.00010
Average[2]	3,568	0.00011[3]

[1] Pumped well.
[2] Wells 4, 4A, 5A, and 6A excluded.
[3] Well 2 excluded.

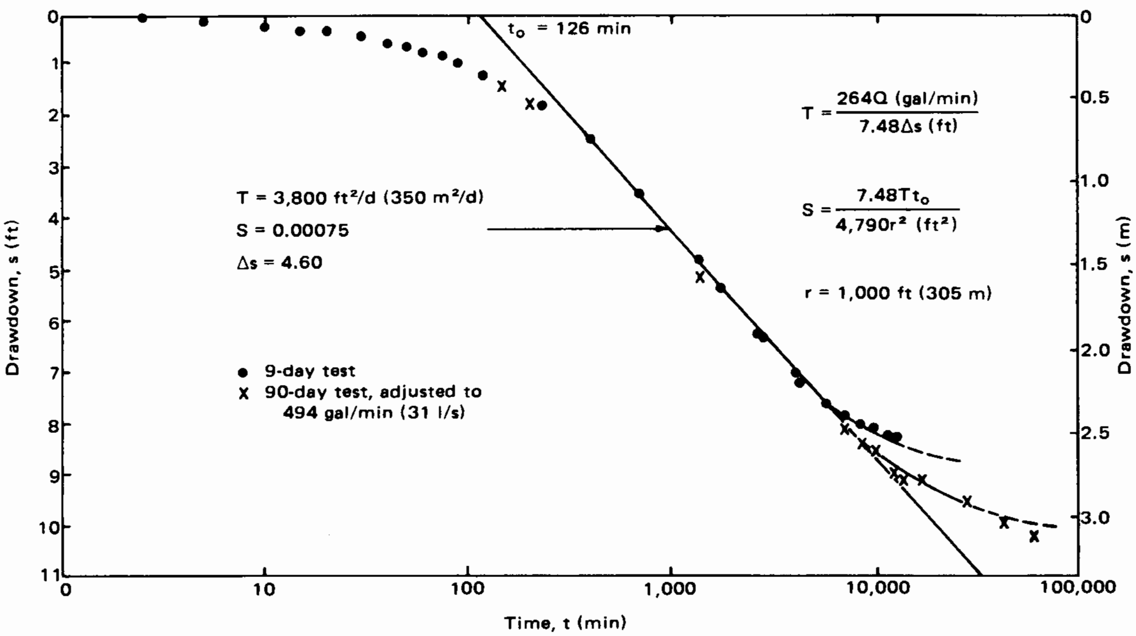

FIGURE 17.—Semilogarithmic graph of drawdown versus time in well 5A during 9-day and 90-day aquifer tests. See text for explanation of equations and symbols.

VOLUMETRIC DETERMINATION OF THE STORAGE COEFFICIENT

The storage coefficient was also determined by a graphical method (fig. 18) in which the volume of water pumped was related to the volume of the cone of depression. The volume of the cone was determined by a computational technique described by Bennett and others (1967, p. 30). Semilogarithmic plots of drawdown versus distance were made for selected time periods, and a straight line was drawn through the plotted points. The volume of the cone was calculated from the formula:

$$V = \frac{\pi}{4.6} \Delta s r_e^2$$

where

V = volume (ft^3),

Δs = line slope (ft/log cycle),

and

r_e = intercept of the straight line with zero drawdown distance (ft).

Because of the poor distribution of the datum points, the position of the straight line in figure 18 is somewhat arbitrary. In calculating the cone volume for the different pumping periods an effort was made to give consistent weight to the individual datum plots. Because relatively slight changes in the r_e distance make large differences in the volume calculation, data collected after 25 days were not considered in this analysis.

In figure 19, volume-of-cone values are plotted against the volume of water pumped, and a line is drawn through the plotted points. Because of the onset of recharge, the line begins to curve upward after about 3 days of pumping.

The storage coefficient was calculated from the slope of a straight line beginning at the origin and tangent to the curve in figure 19. The calculated value, S = 0.00013, is essentially the same as the average value, S = 0.00011, determined for the fully penetrating wells by the time-drawdown straight-line method.

RECHARGE EFFECTS

Recharge to an artesian aquifer can occur in different ways, many of which will produce similar changes either in the slope of semilogarithmic plots of drawdown versus time or in graphs of cone volume versus water pumped. For example, the cone of depression may either spread beyond the confining bed and intersect a surface stream, or recharge may result from general leakage through a semipervious confining bed.

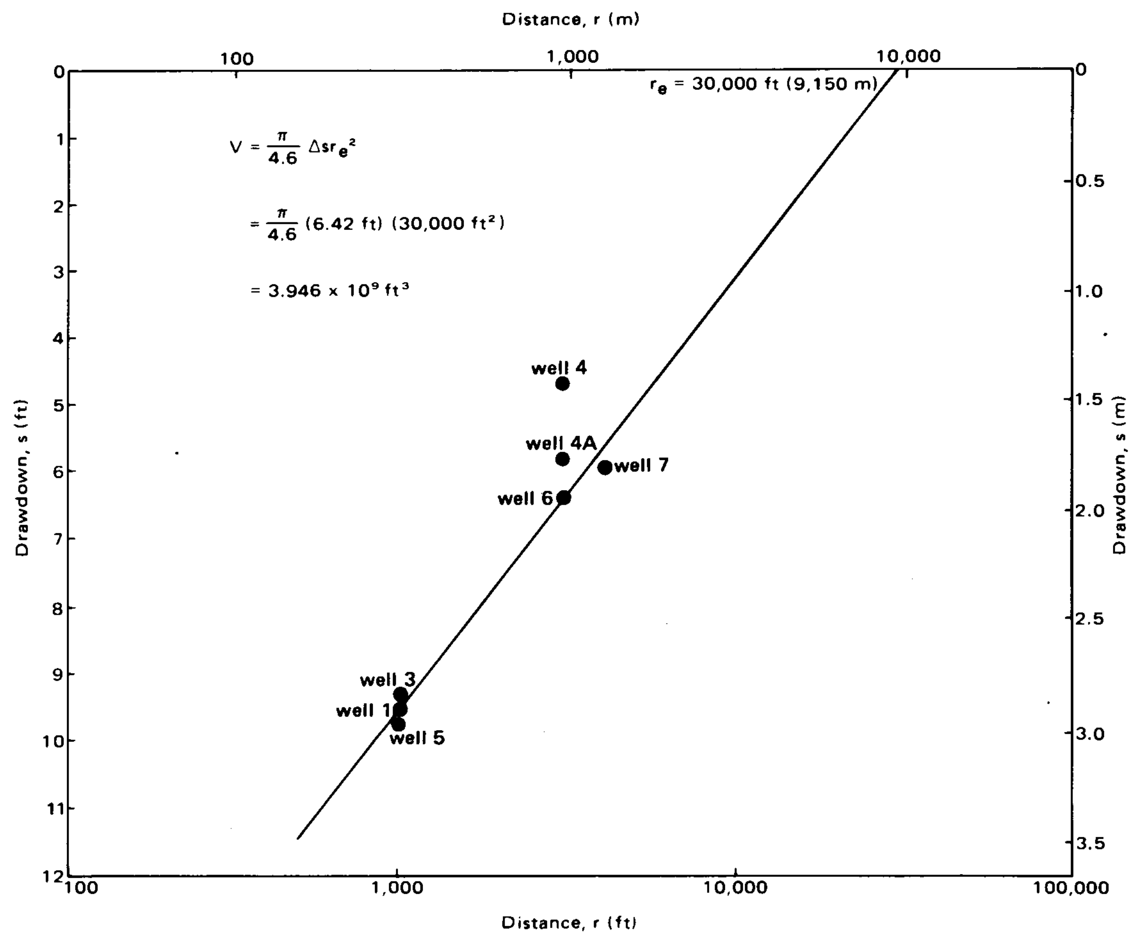

FIGURE 18.—Semilogarithmic graph of drawdown versus distance after 7 days of pumping well 2 at a rate of 369 gal/min (23 l/s). Volume-of-cone determination also shown.

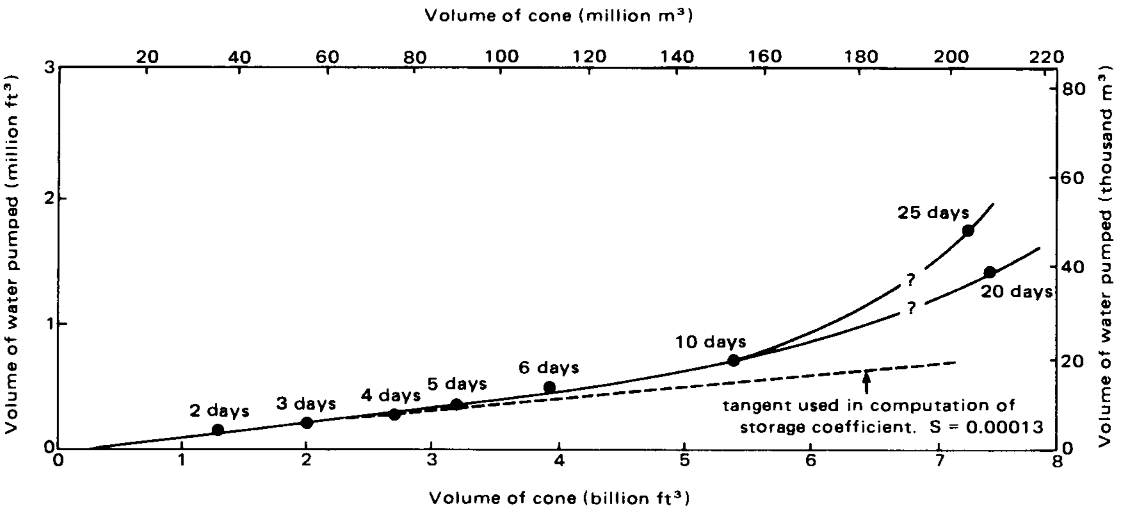

FIGURE 19.—Volume of cone versus volume of water pumped for selected pumping periods. Dashed line is the tangent on which computation of storage coefficient, S, is based.

The Big Island hydrogeologic environment, consisting essentially of a widespread carbonate-rock aquifer overlain by till, is fairly uniform over a much larger area than is affected by the pumping. There is no indication that the aquifer is in free hydraulic communication with the surface at any point within the radius of the cone of depression. The evidence indicates that recharge to the aquifer is from leakage through the overlying regionally extensive mantle of glacial till.

For about half the wells the time-drawdown graphs indicate that recharge effects occurred sooner in the 9-day test than in the 90-day test (see figs. 14, 15, 17). The largest time difference was noted for well 4, in which recharge effects became evident after about 3,500 minutes of pumping in the first test and after about 8,000 minutes in the second test (fig. 14). In wells 1 and 3, on the other hand, recharge effects were evident in both tests after about 4,000 minutes of pumping.

On about half of the graphs the line segment after the break in slope appears straight. The line segment curves upward on the remainder of the graphs, notably those for wells 4 (fig. 14), 4A (fig. 15), 5A (fig. 17), 6, and 7A. Ideally, the lines would be expected to curve upward on all the graphs and eventually approach the horizontal as leakage supplied more and more of the pumpage. The slope of the line would become horizontal (zero slope) when recharge balanced the pumping rate. Equilibrium or near-equilibrium conditions are evident on the semilogarithmic graph for well 4A (see fig. 15). The slope of the line on many of the semilogarithmic graphs does not reflect this condition because inaccuracies may occur in late drawdown data, especially data collected after about 30 days of pumping when frequent rains made accurate drawdown values difficult to determine.

The growth rate of the cone of depression suggests equilibrium conditions did, in fact, prevail in the 90-day test after 20 to 25 days of pumping. Figure 20 shows the change in the apparent effective radius of the cone with time. The graph indicates the cone grew rapidly for the first 10 days of pumping, after which the growth rate was comparatively small. Twenty to 25 days after pumping began the cone was essentially stable. Plots representing 30 or more days of pumping are not shown in figure 20; these plots fall well below the curve, probably because of error in the drawdown data for later times.

Steady-state equation of Jacob

The quantity of water that leaks downward into the aquifer in response to pumping depends on the thickness and vertical hydraulic conductivity of the semipermeable beds through which leakage is occurring, the difference between the water table in the source beds and the potentiometric surface of the aquifer, and the area over which leakage is occurring. The relationship, under steady-state conditions where leakage supplies all the pumpage, is expressed as:

$$Q = (k'/m')\Sigma A \Delta h$$

where

Q = leakage (ft³/d),
k' = hydraulic conductivity (ft/d) of the semipermeable source beds,
m' = thickness (ft) of the beds through which leakage is occurring,
A = area (ft²) over which leakage is occurring,

and

Δh = difference (ft) between the water table and the potentiometric surface.

On the assumption that equilibrium conditions essentially prevailed, the distance-drawdown data for 20 and 25 days of pumping were analyzed using the steady-state "leaky aquifer" equation of Jacob (1946). The equation is given by Ferris and others (1962, p. 110-118) as follows:

$$s = \frac{229QKo(x)}{7.48T}$$

where

s = drawdown (ft) in the observation well,
Q = pumping (leakage) rate (gal/min),
T = transmissivity (ft²/d) of the artesian aquifer,

$Ko(x)$ = modified Bessel function of the second kind of the zero order,

and

x = br/a,

where

r = distance (ft) from the pumped well to the observation well,

a = $(T/S)^{1/2}$, in which S is the storage coefficient (percent) of the artesian aquifer,

and

b = $(k'/m'S)^{1/2}$, in which k' is the hydraulic conductivity (ft/d) of the semipermeable confining bed, and m' is the thickness (ft) of the semipermeable confining bed.

Many of the assumptions inherent in the Theis equation also apply to the Jacob equation. Among these are: the semiconfined aquifer is of infinite areal extent; it is homogeneous and isotropic; the storage coefficient is constant; and water is released instantaneously with a decline in head. It is also assumed that the water level in deposits supplying leakage is constant.

Ferris and others (1962, p. 113) describe a type-curve solution in which the "leaky aquifer" type curve is matched to logarithmic plots of field observations of the drawdown in the observation well, s, and the distance between the pumped well and the observation well, r, at some particular time, t (see Hantush, 1955, and Hantush and Jacob, 1955, for detailed description).

Figure 21 shows the type-curve match to drawdown data collected after 25 days of pumping. Transmissivity based on the type-curve match is 3,300 ft²/d (310 m²/d), about 8 percent lower than the average value of 3,600 ft²/d (330 m²/d), determined from the Theis straight-line analysis of early time-drawdown data. The transmissivity determined from the type-curve match to data collected after 20 days of pumping (not shown) is 3,300 ft²/d (310 m²/d). This close agreement between the transmissivity determinations is further evidence that essentially steady-state conditions prevailed, at least in the region of the pumped well, after about 20 days of pumping.

The leakance factor (k'/m') determined from the Jacob steady-state equation for data collected after 25 days of pumping is 2.9×10^{-6}/d (fig. 21); the value for 20 days of pumping is 3.2×10^{-6}/d.

The leakance factor (k'/m') was also determined from the formula $Q = (k'/m')\Sigma A\Delta h$ in which Δh is the vertical

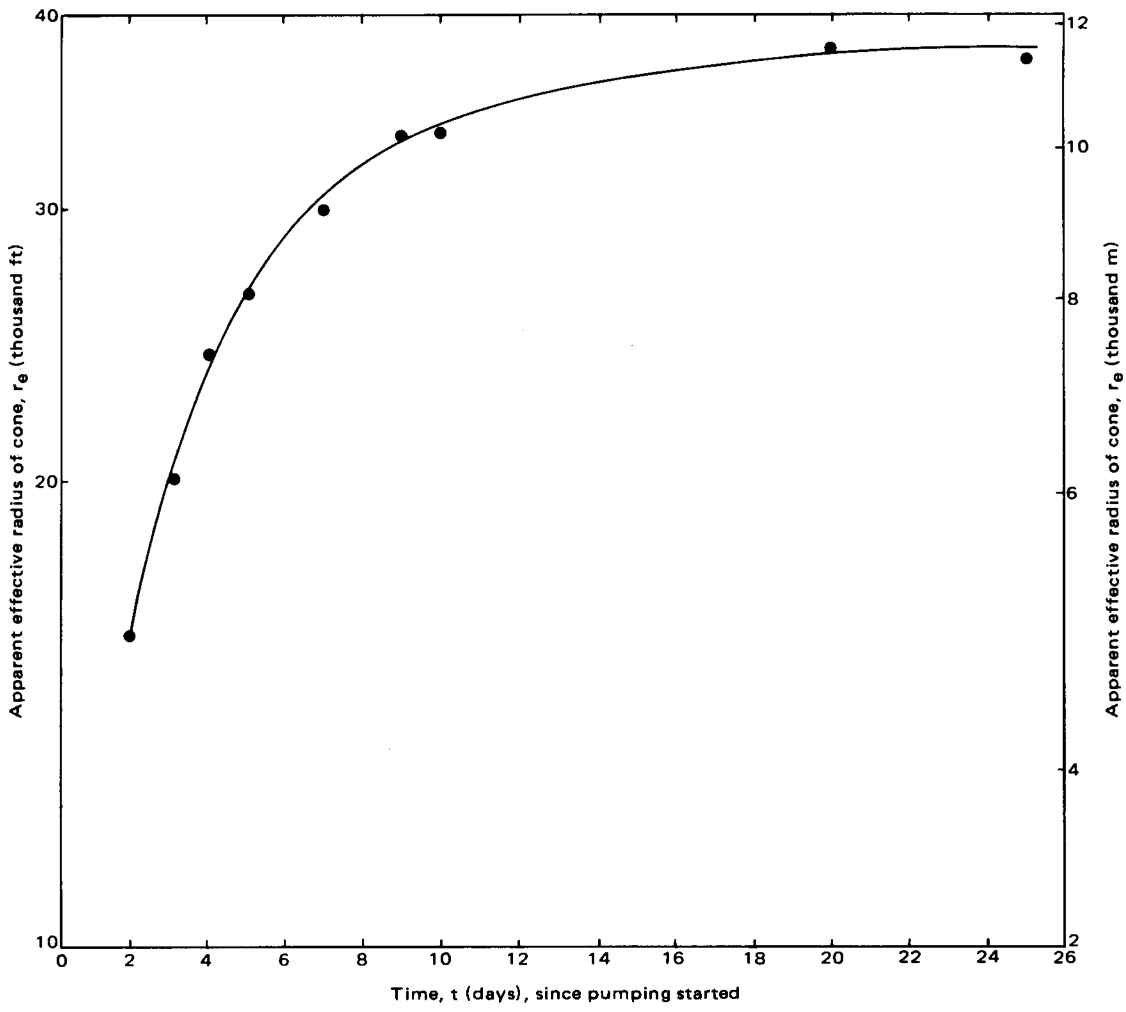

FIGURE 20.–Semilogarithmic graph showing the apparent rate of growth of the cone of depression.

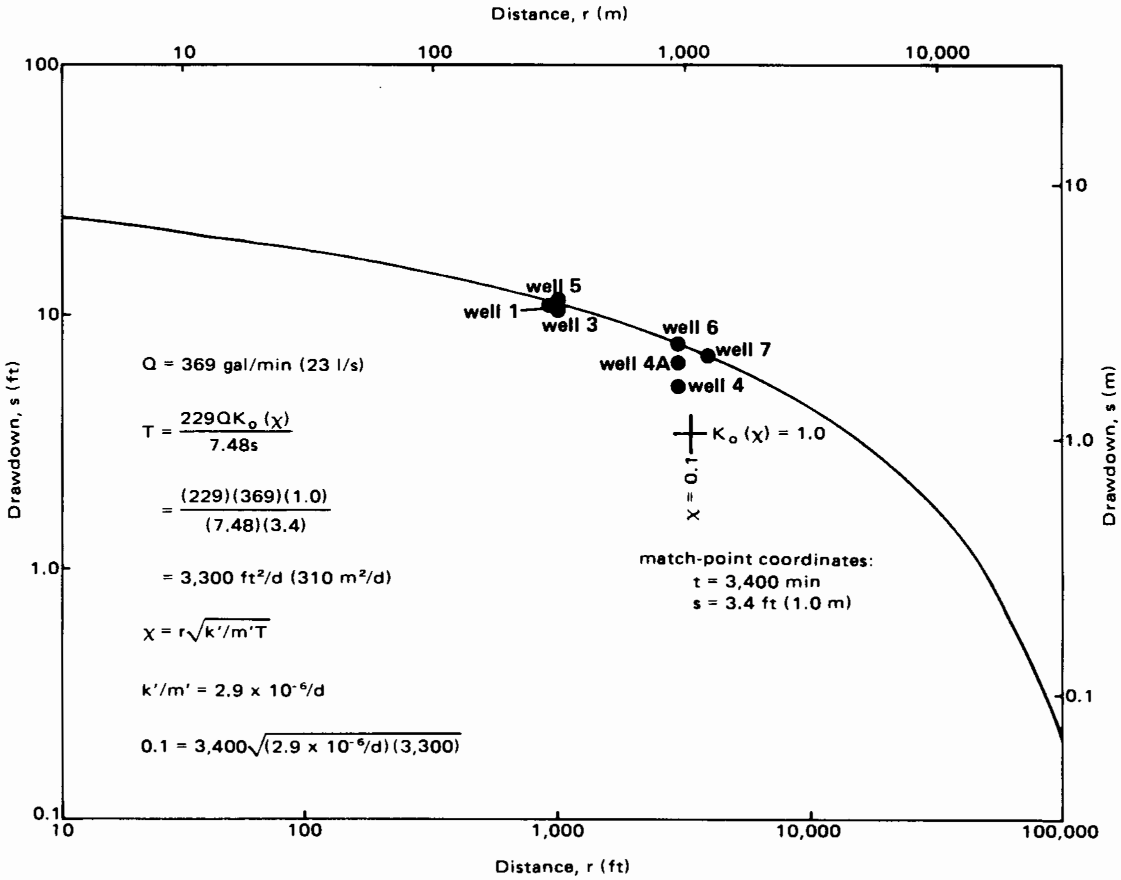

FIGURE 21.—Distance-drawdown logarithmic graph, with leaky aquifer type curve superposed, after 25 days of pumping from well 2. See text for explanation of equations and symbols.

head difference prevailing over an area, A, and the summation of products $A\Delta h$ is taken over the entire area of influence of the well. The term $A\Delta h$ may be evaluated as the volume of the cone of depression at a time when all the pumped water is assumed to come from leakage. The volume of the cone determined by the method illustrated in figure 18 for 20 and 25 days after pumping started was 7.397×10^9 and 7.240×10^9 ft³ (2.09×10^8 and 2.05×10^8 m³), respectively. Dividing the daily pumping rate by each of these cone volumes yields values for the leakance factor of 9.6×10^{-6}/d and 9.8×10^{-6}/d for 20 and 25 days, respectively. These values compare reasonably well with values determined from the steady-state type-curve method of Jacob for the same periods of pumping, 3.2×10^{-6}/d and 2.9×10^{-6}/d, respectively. Because both methods of analysis independently yield values for the leakage coefficient in the same order of magnitude, the assumption of steady-state conditions prevailing after about 20 days of pumping is probably justified.

Vertical hydraulic conductivity of the till

The semiconfining bed through which leakage is occurring to the aquifer at Big Island and which also constitutes the source bed consists of dense claylike till. Based on the logs of the test wells and logs of about 25 farm and domestic wells, the average thickness of the till is about 31 ft (9 m) at the test site and about 44 ft (13 m) in the general area. The range in till thickness recorded in these logs is 21 to 89 ft (6 to 27 m). The hydraulic conductivity of the till, based on the range of values determined for the leakance factor (2.9×10^{-6}/d to 9.8×10^{-6}/d) and its estimated average regional saturated thickness of 35 ft (11 m), is in the range of 1.0×10^{-4} to 3.4×10^{-4} ft/d (3.0×10^{-5} to 1.0×10^{-4} m/d).

This range is low compared with hydraulic conductivity values listed by Norris (1963) for till at sites in Ohio, Illinois, and South Dakota. Those values, some of which were determined by aquifer tests and others from permeameter tests made in the laboratory, range from a low of 4.0×10^{-5} ft/d (1.2×10^{-5} m/d) to a high of 0.12 ft/d (0.04 m/d), with most values falling in the range of 1.0×10^{-3} to 4.0×10^{-2} ft/d (3.0×10^{-4} to 1.2×10^{-2} m/d). Thus, the vertical hydraulic conductivity of the till at Big Island is about an order of magnitude less than that of most of the tills listed by Norris. However, despite the relatively low conductivity of the till at Big Island, a large amount of water is available to wells from vertical leakage because of the large area over which the cone of depression can develop.

It is interesting to consider the amount of till dewatering required to yield the volume of water pumped, 1.4×10^6 ft³ (3.9×10^4 m³), during the first 20 days of the aquifer test. Assuming a cone radius of 35,000 ft (10,700 m) and a specific yield of 20 percent, the thickness of dewatered till required to yield this volume of water would average about 1.8 x 10^{-3} ft (5.5 x 10^{-4} m).

To cause such dewatering to occur a head difference would have to be maintained between the water table in the till and the potentiometric surface in the carbonate-rock aquifer. For a cone with a radius of 35,000 ft (10,700 m), using the steady-state formula $Q = (k'/m')\Sigma A \Delta h$, the average head differential required to support a pumping rate of 369 gal/min (23 l/s) would be about 6 ft (2 m) for a leakance factor of 2.9 x 10^{-6}/d. It is evident that the till as a source bed, despite its relatively low hydraulic conductivity, is capable of yielding a very large quantity of water to the underlying aquifer when a large area of leakage is considered and a relatively large head difference is induced.

SUMMARY OF AQUIFER-TEST RESULTS

Profiles of the cone of depression and specific-capacity tests of individual wells indicate that the carbonate-rock aquifer near Marion, Ohio, is heterogeneous, with hydraulic conductivity being highest near the center of the test-well array and lowest in the vicinity of well 4. An analysis of semilogarithmic distance-drawdown graphs, supplemented by a time-drawdown type-curve analysis of data from a special test of well 2, indicates that transmissivity near wells 2 and 3 is in the range of 50,000 to 70,000 ft²/d (4,600 to 6,500 m²/d).

Distance-drawdown graphs for selected pumping periods show that the shape of the cone of depression became essentially stable after a day or two of pumping. The profile of the cone was not significantly different after 20 days of pumping than it was after 2 days of pumping. This suggests that tests of similar aquifers need not be of long duration to yield essential hydrologic data.

Transmissivity determined for the outer part of the cone from a 7-day drawdown-contour map was about 2,800 ft²/d (260 m²/d). Although this value is low compared with those determined by other techniques for the area closer to the pumped well, it is probably representative of the average regional character of the aquifer.

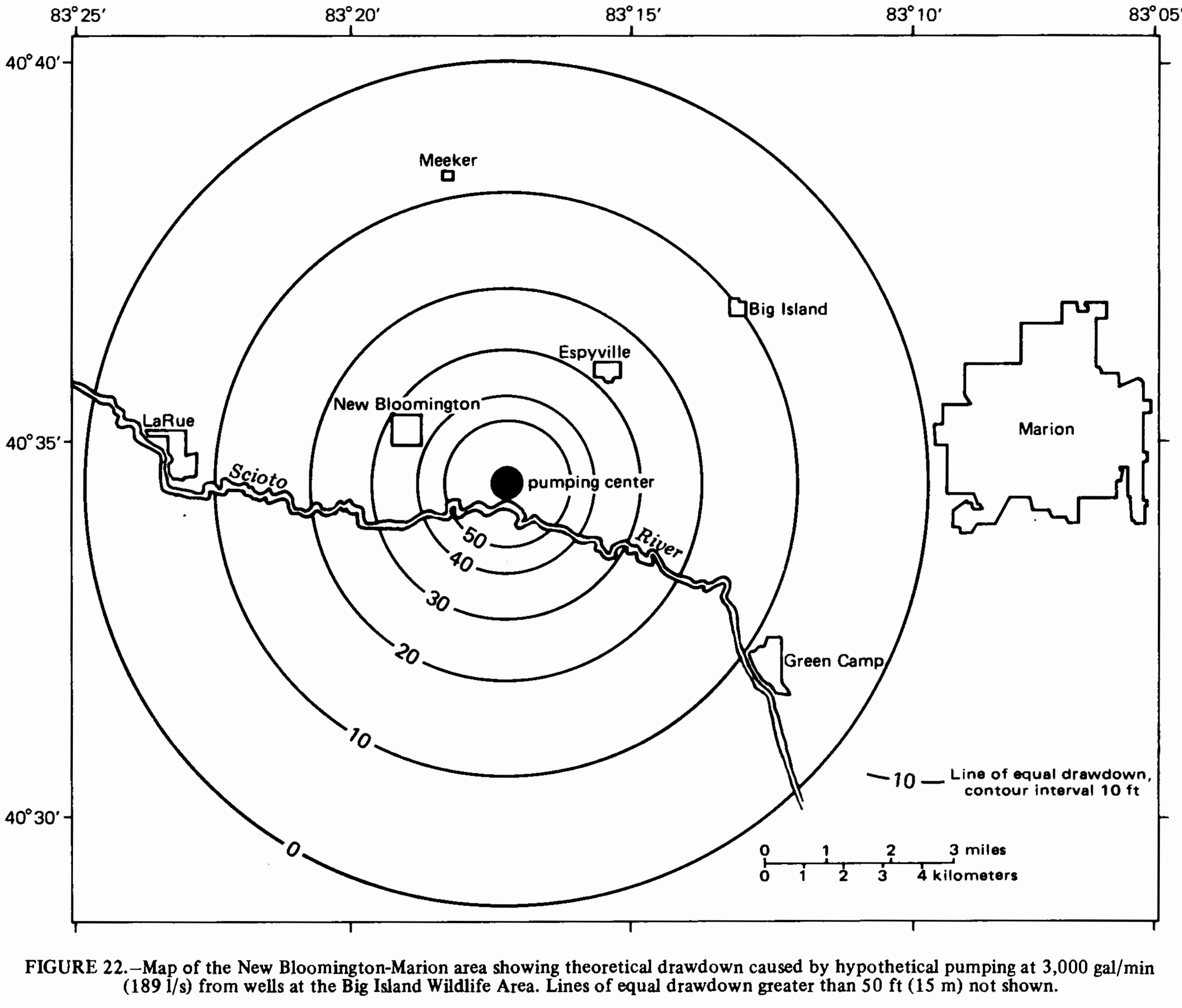

FIGURE 22.—Map of the New Bloomington-Marion area showing theoretical drawdown caused by hypothetical pumping at 3,000 gal/min (189 l/s) from wells at the Big Island Wildlife Area. Lines of equal drawdown greater than 50 ft (15 m) not shown.

Transmissivity determined from semilogarithmic time-drawdown data from the fully penetrating test wells averaged 3,600 ft^2/d (330 m^2/d). The drawdown data used in this analysis were from the test interval between 200 to 300 minutes and 4,000 to 6,000 minutes after pumping began, or the period when the data plots began to fall on a straight line until the onset of recharge, which slowed and eventually stopped the growth of the cone. The time-drawdown analysis is believed to have essentially averaged the effects of differences in hydraulic conductivities between regions and yielded an integrated value for transmissivity.

The storage coefficient determined from early time-drawdown data collected before recharge began to affect development of the cone was 10^{-4}, typical of artesian aquifers. The evidence indicates that most of the water released from storage in the aquifer was from the area beyond the region of high hydraulic conductivity. The highly conductive region, which is small compared with the aquifer as a whole, acts primarily as a zone of transmission rather than as a zone of storage release.

An arithmetic plot relating the volume of water pumped to the volume of the cone of depression (fig. 18) departs from a straight line after about 3 days of pumping, indicating the effect of recharge. The same effect was evident on the semilogarithmic time-drawdown graphs, causing an upward break in the line slope after 4,000 to 6,000 minutes of pumping. A graph (fig. 20) showing the rate of growth of the cone indicates that essentially steady-state conditions prevailed after about 20 to 25 days of pumping. The hydrologic evidence indicates that recharge to the aquifer is from vertical leakage through the overlying till.

The leakance factor, k'/m', where k' is the vertical hydraulic conductivity of the semiconfining till bed and m' is its saturated thickness, was determined by a type-curve method based on the Jacob steady-state "leaky-aquifer" equation and by a volumetric method. The values were in the same order of magnitude. For the type-curve method they were 3.2×10^{-6}/d for a pumping period of 20 days and 2.9×10^{-6}/d for a pumping period of 25 days; values determined by the volumetric method for the same periods were 9.6×10^{-6}/d and 9.8×10^{-6}/d, respectively.

Based on the estimated average regional saturated thickness of 35 ft (11 m), the vertical hydraulic conductivity of the till is 1.0×10^{-4} to 3.4×10^{-4} ft/d (3.0×10^{-5} to 1.0×10^{-4} m/d). This range is lower than some previously published (Norris, 1963) values for till at sites in other parts of Ohio and in Illinois and South Dakota.

Despite its relatively low hydraulic conductivity, calculations show that the till can supply large quantities of water to the aquifer with minimal decline in saturated thickness when areas comparable in size to the area affected by the aquifer test are considered and sufficient head difference is induced between the water table in the till and the potentiometric surface in the underlying aquifer.

PRACTICAL CONSIDERATIONS IN THE USE OF GROUND WATER AT BIG ISLAND

The most important factor to consider in large-scale use of ground water at the Big Island Wildlife Area is the effect of pumping on ground-water levels in the surrounding area. Because ground-water levels generally are within a few feet of the land surface, many farm and domestic wells are equipped with shallow-well suction pumps. The lowering of ground-water levels by only a few feet in much of the area would require either the lowering of pump intakes, replacement of pumps, or the redrilling or deepening of wells in some instances.

During the aquifer test the drawdown in well 7, located 4,000 ft (1,219 m) from the pumped well, amounted to about 7 ft (2 m) after 25 days of pumping. Yet the pumping rate during the test, 369 gal/min (23 l/s), was only about one-eighth as high as the normal withdrawal rate from the river to maintain the wildfowl habitat area.

If wells 1, 2, and 3 together could be made to yield 3,000 gal/min (189 l/s), the rate now being pumped from the river and the rate originally planned when the wells were drilled, the drawdown would be substantial over a very large area. Figure 22 shows the probable magnitude and distribution of the drawdown based on the observed drawdowns in the test wells adjusted to a pumping rate of 3,000 gal/min (189 l/s) after 25 days of pumping. The limit of the cone of depression is assumed to be 35,000 ft (10,700 m).

The real lines of equal drawdown, were pumping to become an actuality, would not be as symmetrical as those shown because of the heterogeneity of the aquifer. The shape of the cone of depression would more nearly resemble the configuration in figure 12, which is based on drawdowns measured in the field after 7 days of pumping from well 2.

As figure 22 shows, a drawdown of about 10 ft (3 m) could be expected at Big Island, 4.5 mi (7 km) from the center of pumping, and a drawdown of at least a few feet would occur as far away as LaRue and Green Camp, about 5 mi (8 km) from the center of pumping. More importantly, drawdowns of between 30 and 40 ft (9 and 12 m) would occur at the village of New Bloomington, which has no public water-supply system and whose 350 residents rely on individual wells.

If the social and economic problems associated with a substantial regional lowering of ground-water levels could be solved, there would still remain the practical question of how much water can be pumped from the existing three wells. As figure 22 shows, drawdown increases rapidly toward the center of pumping. Although wells 1, 2, and 3 are fortuitously located within a highly transmissive zone and have relatively high specific capacities, well-interference effects would be substantial if these wells were pumped simultaneously. Not considering the effect of aquifer dewatering, the estimated drawdown in well 2 with wells 1, 2, and 3 all being pumped at the rate of 1,000 gal/min (63 l/s) is approximately 90 ft (27 m). This includes formational drawdown, well loss, and interference effects.

The estimated drawdown in wells 1 and 3 under the same pumping regime is about 120 ft (37 m). Drawdown in these wells would be greater than in well 2 despite the fact that wells 1 and 3 are the end wells in the three-well alignment. Well 2 has the highest specific capacity of all the test wells (see table 2). Well losses also are low in well 2.

Relative to well losses, if permanent pumps were installed in wells 1 and 3, these wells probably would be enlarged as well 2 was. Enlarging the wells might improve their efficiencies somewhat and lessen the component of drawdown due to turbulent flow. This happened to well 2 when it was enlarged (Norris, 1976). Such an improvement in efficiencies of wells 1 and 3 probably would not cause a significant change in their pumping levels, however.

An important additional factor in estimating pumping levels is aquifer dewatering. Drawdowns in the range of 90 to 120 ft (27 to 37 m) would mean that dewatering in the vicinity of the wells would be 55 to 80 percent of the aquifer thickness, assuming that ground-water levels were about 5 ft (2 m) below the surface before pumping started. Dewatering of this magnitude with its attendant reduction in transmissivity would increase the drawdown by an estimated 40 to 50 percent. This means that pumping levels in the wells would be near the base of the aquifer. Complete dewatering of the aquifer is impossible from a practical standpoint, but it is evident that a pumping rate of 3,000 gal/min (189 l/s) could not be sustained.

It appears to the author that a pumping rate of about 500 gal/min (32 l/s) from each of the three wells would remain a practical possibility, however. The regional effects of pumping at this reduced rate can readily be determined from figure 22 by dividing each of the drawdown contours by two.

Problems involving the impact of pumping on farm and domestic wells in the area would still exist at the reduced withdrawal rate, but such problems would be much less severe and possibly could be solved on an individual basis. If ground water could be used to supply half the water needed at the wildlife area, perhaps there would be less concern over pumping the remainder from the Scioto River.

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